Saskatchewan
Energy and Mines

Saskatchewan
Geological Survey

Miscellaneous Report 91-4

Summary of Investigations 1991, Saskatchewan Geological Survey

Printed under the authority of the
Minister of Energy and Mines

Although the Department of Energy and Mines has exercised all reasonable care in the compilation, interpretation and production of this report, it is not possible to ensure total accuracy, and all persons who rely on the information contained herein do so at their own risk. The Department of Energy and Mines and the Government of Saskatchewan do not accept liability for any errors, omissions or inaccuracies that may be included in, or derived from, this report.

Editors

R. Macdonald
T.I.I. Sibbald
C.T. Harper
D.F. Paterson
P. Guliov

Production Editors:

C.T. Harper
C.L. Brown

Production Assistants:

J. Bilkhu
V. Zeigler
R.P. Kydd
K.G. Jones

ISSN 0228 5657
Printed by Printco Graphics Inc.

Foreword

This summary report presents preliminary results of the geoscience program of Saskatchewan Energy and Mines conducted by staff of the Saskatchewan Geological Survey and its contractors. Also included are a number of contributions from outside agencies including the Geological Survey of Canada, Saskatchewan Research Council and the geology departments of the Universities of Regina and Saskatchewan.

Operational funds of the Saskatchewan Geological Survey are provided largely through a new five-year Canada-Saskatchewan Partnership Agreement on Mineral Development (known as "MDA2") which was signed in May 1991, retroactive to April 1, 1990. Expenditures under the 1990-91 Saskatchewan Mineral Industry Diversification Program are eligible as the provincial contribution in year one of the agreement. The federal part of the program is being implemented in the four years 1991-95, with provision for wind up and final reporting in fiscal year 1995-96. Total funding for the five-year program is $10 million, equally cost-shared between Canada and Saskatchewan on a parallel delivery basis. MDA2 is primarily oriented to geoscience with a total of $7.35 million allocated to geological mapping, geophysical and geochemical surveys, mineral deposit investigations and related studies. Other components of the agreement include research and development in mineral processing, market and other economic studies, and public information.

In 1991-92, approximately 60 percent of provincial MDA2 funds were directed to northern geoscience field projects, 20 percent to southern geoscience and the remainder to compilation and data-base projects as well as prospector training and public information.

In the northern program, emphasis switched from the La Ronge Belt to the southeastern part of the Shield around Flin Flon, Hanson Lake and portions of the Glennie and Kisseynew Domains. In this region, revision geological mapping and geochemical studies are being carried out in an attempt to gain a better understanding of the geological setting and control of gold and base-metal deposits, as well as to identify areas worthy of detailed exploration.

An important feature of the 1991 geoscience program is a higher degree of integration of federal and provincial projects than had been attempted in previous federal-provincial parallel delivery programs. For example, provincial projects in the Flin Flon-Hanson Lake area (Project Seagull) form a component of the broader Geological Survey of Canada's Shield Margin Project of the National Mapping Program (NatMap). This program involves data acquisition in digital format and GIS-generated bedrock geology maps, and will serve as a useful test case for further development of computerized mapping in the provincial survey. The multidisciplinary nature of the Shield Margin Project is being enhanced by the Lithoprobe Transect of the Trans-Hudson Orogen which crosses the Project Seagull area.

In the south, a number of projects started under MDA1 and the Saskatchewan Mineral Industry Diversification Program are being continued under MDA2 funding. These include studies on brines and building stones, lignite reserve estimates, and an investigation of the regional distribution of kimberlites and their geophysical and geochemical signatures. During the year the Geological Survey of Canada, Saskatchewan Geological Survey and several industry members initiated a joint program of aeromagnetic surveys intended to complete coverage of the Sedimentary Basin in four years. The new aeromagnetic data will be invaluable for kimberlite prospecting as well as for studies on potash and hydrocarbons.

I wish to thank all of the authors who contributed papers to Summary of Investigations 1991 and gratefully acknowledge the diligence of the editorial, cartographic and stenographic staff in maintaining the high standard of publication.

L.S. Beck
Executive Director
Geology and Mines Division

October 1991

Contents

Page

Foreword iii

Contents v

Location Map, Northern Projects viii

A. Northern Projects Under the Canada-Saskatchewan Partnership Agreement on Mineral Development 1990-95, in Year 1991-92

* Revision Bedrock Geological Mapping, Bootleg Lake-Birch Lake Area (Parts of NTS 63K-12 and 63L-9) .... *D.J. Thomas* 2

* Revision Bedrock Geological Mapping, Mystic-Kaminis Lakes Area (Parts of NTS 63K-12 and 63L-9) .... *B.A. Reilly* 9

* Revision Bedrock Geological Mapping, Table Lake Area (Part of NTS 63L-09) .... *W.L. Slimmon* 16

Lithogeochemical Studies, East Amisk Lake Area .... *B.R. Watters* 21

The NatMap Shield Margin Project and Project Seagull in Saskatchewan: An Integrated Geoscience Program Incorporating GIS .... *R. Macdonald, S. Lucas, D.J. Thomas and B. Czornobay* 28

* Revision Bedrock Geological Mapping, Wildnest-Attitti Lakes Area (Parts of NTS 63M-1 and -2) .... *K. Ashton and A.D. Leclair* 29

Metamorphic P-T Investigations in the Southern Part of Kisseynew and Western Part of Flin Flon Domains .... *S. Diegel, K. Ashton and K. Wilcox* 41

* Revision Bedrock Geological Mapping, Deschambault-Oskikebuk Lakes Area (Parts of NTS 73I-16, 63L-11, -12, -13 and -14) .... *G.D. Delaney* 47

* Revision Bedrock Geological Mapping, Southeast Arm, Deshambault Lake-Northern Lights Area (Part of NTS 63L-10, -11 and -14) .... *J.F. Lewry, T.I.I. Sibbald and N. Hayward* 57

Glennie-Kisseynew P-T Study 1991: Fieldwork and Preliminary Results .... *D. Perkins* 66

Metamorphic P-T Studies in the Jan Lake Area, Hanson Lake Block .... *S. Abbas-Hasanie, J.F. Lewry and D. Perkins* 69

Lithogeochemical Studies, Wildnest-Attitti Lakes Area .... *G.R. Parslow* 72

U-Pb Geochronological Investigations in the Trans-Hudson Orogen, Saskatchewan: Preliminary Results by the ROM Laboratory in 1990-91 .... *L.M. Heaman, S.L. Kamo, G.D. Delaney, C.T. Harper, B.A. Reilly, W.L. Slimmon and D.J. Thomas* 74

The Age of the Mine Granites, Goldfields Area .... *D. O'Hanley, T.K. Kyser and T.I. Sibbald* 76

Deformation in the Star Lake-McLennan Lake Area, La Ronge Domain .... *R.G. Roberts and He Chei* 81

Alteration Halos Associated with Gold-bearing Shear Zones in the Star Lake and Island Lake Plutons .... *E.C. Appleyard* 85

Field Observations on the Broken Hammer Gold Occurrence and NIKO Gold Deposit .... *P. Schwann* 93

Structural Controls and Geochemical Character of the Seabee Gold Mine, Laonil Lake, Glennie Domain .... *D.J. Schultz and R. Kerrich* 101

Mineralization and Associated Alteration in the Hanson Lake Area .... *R.O. Maxeiner and B.R. Watters* 109

* Rare Element Pegmatites in the Hanson Lake Pegmatite Field .... *D.G. MacDougall* 118

Mineralized Core Collection .... *A.J. Gracie* 127

B. Northern Projects Under the Canada-Saskatchewan Partnership Agreement on Mineral Development 1990-95, from Year 1990-91

Geochemistry and Tectonic Setting of Metabasaltic Rocks from the Snake Rapids Area,
Flin Flon Domain .... *B.R. Watters and K.E. Ashton* 130

Geological Relationships in the Wood Lake Area,
Tabbernor Fault Zone, Saskatchewan .... *K.H. Wilcox* 135

A Reconnaissance Petrographic and Isotopic Study of Fluid History
in the Tabbernor Fault Zone, Saskatchewan .... *M.P. Field and R. Kerrich* 144

C. Contributed Papers, Northern Studies

The Lithoprobe Trans-Hudson Orogen Transect (THOT): Status Report .... *J.F. Lewry and Z. Hajnal* 154

Detrital Zircons from Missi Metasedimentary Rocks, Flin Flon Basin: Constraints on
Age and Provenance .... *K.M. Ansdell, T.K. Kyser, M.R. Stauffer and G. Edwards* 157

Chemical Signatures of Igneous and Metasedimentary Rocks from the Pelican Slide Area:
Implications for their Sources and Tectonic environments
.... *M.Sun, M.R. Stauffer, J.F. Lewry, G. Edwards, R. Kerrich and T.K. Kyser* 162

* Origin and Kinematic History of Highly Strained Gneisses in the
Jan Lake East Area .... *Shi Rupan and J.F. Lewry* 169

Petrography, Fluid Inclusion and Stable Isotope Data from the Bakos Lode Gold Deposit,
La Ronge Domain .... *M. Fayek, T.K. Kyser and R.T. Kusmirski* 175

Mineral Parageneses, Fluid Characteristics and Radiogenic Isotopic Data from the Proterozoic
Jasper Gold Zone, La Ronge Domain .... *F. Hrdy, T.K. Kyser and R.T, Kusmirski* 178

Tourmaline, Phosphate Minerals, Zircon and Pitchblende in the Athabasca Group:
Maw Zone and McArthur River areas, Saskatchewan .... *D. Quirt, T. Kotzer and T.K. Kyser* 181

The East Athabasca Mylonite Zone: An Archean Segment of the Snowbird Tectonic Zone
in Northern Saskatchewan .... *S. Hanmer, C. Kopf and M. Darrach* 192

D. Southern Projects Under the Canada-Saskatchewan Partnership Agreement on Mineral Development 1990-95, in Year 1991-92

Review of Industrial Minerals Investigations 1991-92 .... *P. Guliov and M. Gent* 198

Industry-Government Aeromagnetic Survey
of Southern Saskatchewan .... *D. Teskey, M.R. Gent and P.E. Stone* 201

E. Other Southern Studies

Projects Supported by Petroleum Geology Branch 204

Note on the Ordovician-Silurian Boundary in Southeastern Saskatchewan .... *F.M. Haidl* 205

Preliminary Palynological Investigations of Upper Cretaceous Strata,
Southwestern Saskatchewan .... *C.F. Gilboy and E.M.V. Nambudiri* 208

F. Appendices

Titles and Abstracts of Recent Saskatchewan Energy and Mines Publications and Papers by Staff of the Saskatchewan Geological Survey 212

Acknowledgments 219

Organization Chart 222

Staff List .... *inside back cover*

---

* Asterisked reports are accompanied by maps, as listed in full on the accompanying map package.

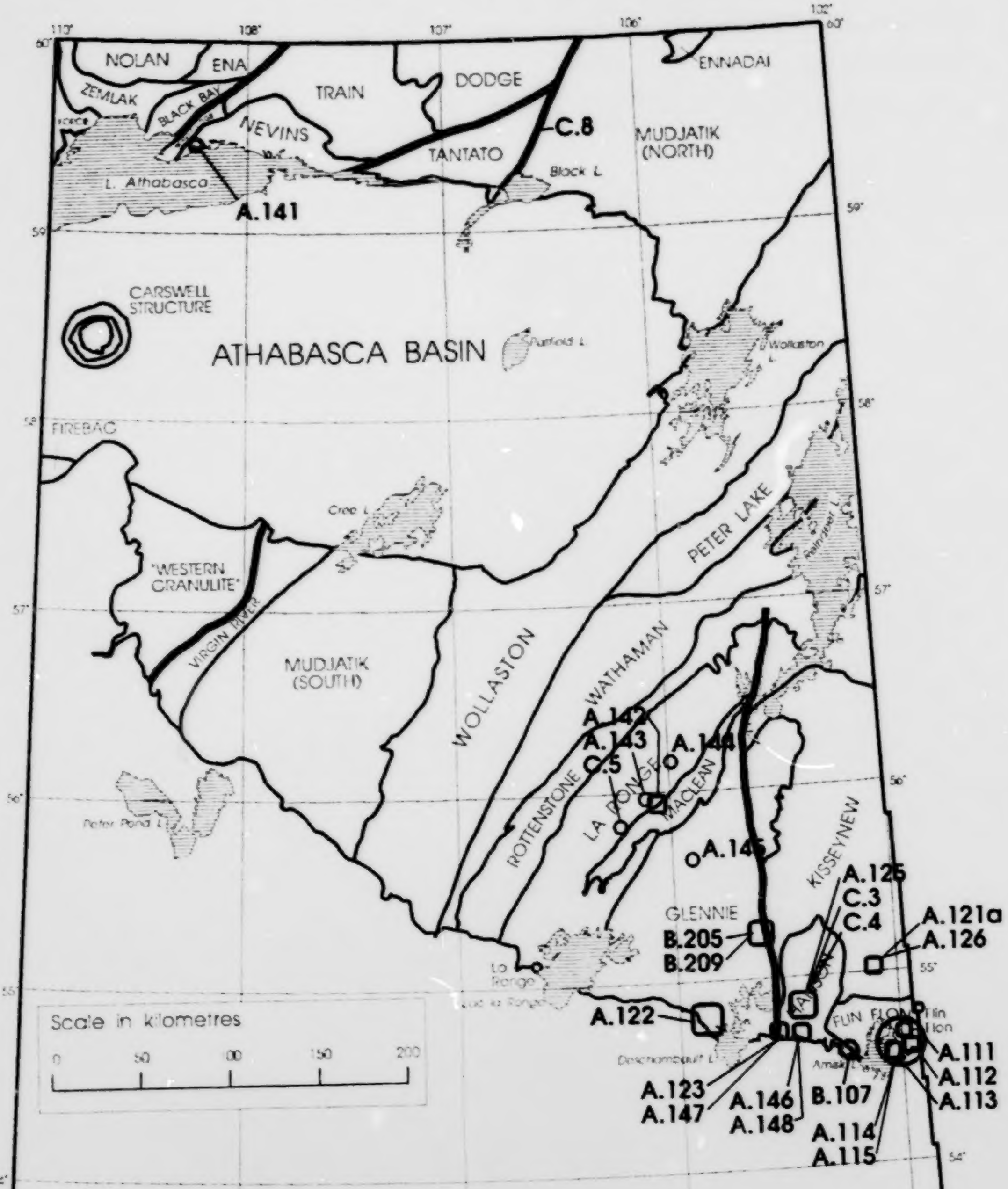

110°
108°
107°
106°
102°
60°
NOLAN
ENA
ZEMLAK
BLACK BAY
TRAIN
DODGE
ENNADAI
NEVINS
C.8
MUDJATIK (NORTH)
TANTATO
Black L.
L. Athabasca
A.141
59°
CARSWELL STRUCTURE
ATHABASCA BASIN
Pasfield L.
Wollaston L.
58°
FIREBAG
Cree L.
PETER LAKE
Reindeer L.
"WESTERN GRANULITE"
VIRGIN RIVER
57°
MUDJATIK (SOUTH)
WOLLASTON
WATHAMAN
A.142
A.143
A.144
C.5
LA RONGE
MACLEAN
ROTTENSTONE
KISSEYNEW
A.145
56°
Peter Pond L.
A.125
C.3
C.4
A.121a
A.126
GLENNIE
B.205
B.209
La Ronge
Lac la Ronge
55°
A.122
Scale in kilometres
0
50
100
150
200
FLIN FLON
Flin Flon
Deschambault L.
Amisk L.
A.123
A.147
A.146
A.148
B.107
A.111
A.112
A.113
A.114
A.115
54°

Figure 1 - Location of projects or reports in hand 1991-92 in northern Saskatchewan (excluding compilation maps). Numbers on map refer to the project number of the Canada-Saskatchewan Partnership Agreement on Mineral Development program (PAMD or generally known as MDA2) listed in the first column opposite. The base map shows in abbreviated form the location of the main lithostructural domains in northern Saskatchewan.

Outputs from PAMD 1990-5 Projects in 1991-92:

Proj. No.	Data No.*	Short Title of Paper or Project *Author(s)*	
A.111	9112	Revision Bedrock Mapping, Bootleg Lake-Birch Lake Area *Thomas, D.J.*	(F,M)
A.112	9123	Revision Bedrock Mapping, Mystic-Kaminis Lakes Area *Reilly, B.A.*	(F,M)
A.113	9130	Revision Bedrock Mapping, Table Lake Area *Slimmon, W.L.*	(F,M)
A.114	9137	Lithogeochemical Studies, East Amisk Lake Area *Watters, B.R.*	(F,C)
A.115	9150	Shield Margin Project (NatMap), Project Seagull and GIS *Macdonald, R. et al.*	(G)
A.121a	9147	Revision Bedrock Mapping, Wildnest-Attitti Lakes Area *Ashton, K.E. & Leclair, A.D.*	(F,M)
A.121b	9147	Metamorphic P-T Investigations, Kisseynew and Flin Flon Domains *Diegel, S. et al.*	(W,C)
A.122	9122	Revision Bedrock Mapping, Deschambault-Oskikebuk Lakes Area *Delaney, G.D.*	(F,M)
A.123 A.147	9117 9106	Revision Bedrock Mapping, Southeast Arm, Deschambault Lake-Northern Lights Area *Lewry, J.F. et al.*	(F,M,C)
A.124	9136	Glennie-Kisseynew P-T Study 1991 *Perkins, D.*	(F,C,W)
A.125	9151	Metamorphic P-T Studies, Jan Lake Area *Abbas-Hasanie, S. et al.*	(F,C)
A.126	9115	Lithogeochemical Studies, Wildnest-Attiti Lakes Area *Parslow, G.R.*	(F,C)
A.131	9128	U-Pb Geochronological Investigations in Northern Saskatchewan: Preliminary Results by ROM in 1991-92 *Heaman, L.M. et al.*	(F,C,W)
A.132	9104	Review of Recent Isotopic/Geochronological Work *Bickford, M.E. et al.*	*(F,C,W,N)*
A.141	9134	Age of the Mine Granites, Goldfields Area *O'Hanley, D. et al.*	(F?,C)
A.142	9129	Deformation in the Star-McLennan Lakes Area *Jie, H. & Roberts, R.G.*	(F,C)
A.143	9138	Alteration Halos, Star and Island Lake Plutons *Appleyard, E.C.*	(F,C)
A.144	9135	Field Observations of the Broken Hammer Gold Showing and the NIKO Gold Deposit *Schwann, P.*	(F)
A.145	9149	Structural Controls and Geochemical Character, Seabee Gold Mine *Schultz, D.J. et al.*	(F?,C)
A.146	9152	Mineralization and Associated Alteration in the Hanson Lake Area *Maxeiner, R.O. & Watters, B.R.*	(F,C)
A.148	9133	Rare Element Pegmatites in the Hanson Lake Pegmatite Field *MacDougall, D.G.*	(F,M)
A.321	9105	Mineralized Core Collection *Gracie, A.J.*	(W)

Outputs from PAMD 1990-5 in 1990-91 and extension:

Proj. No.	Data No.*	Short Title of Paper or Project *Author(s)*	
B.107	9037	Geochemistry and Tectonic Setting of Metabasaltic Rocks from the Snake Rapids Area *Watters, B.R. & Ashton, K.*	(C)
B.205	9040	Geological Relationships in the Wood Lake Area, Tabbernor Fault Zone *Wilcox, K.H.*	(F,G)
B.209	9048	Reconnaissance Petrographic and Isotopic Study of the Fluid History in the Tabbernor Fault Zone *Field, M.P. & Kerrich, R.*	(C)

Contributed Papers:

Proj. No.	Data No.*	Short Title of Paper or Project *Author(s)*	
C.1		The Lithoprobe Trans-Hudson Transect (THOT): Status Report *Lewry, J.F. & Hajnal Z.*	(W)
C.2		Detrital Zircons from Missi Metasedimentary Rocks, Flin Flon Basin: Constraints on Age and Provenance *Ansdell, K.M. et al.*	
C.3		Chemical Signatures of Igneous and Metasedimentary Rocks from the Pelican Slide Area: Implications for their Sources and Tectonic environments *Sun, M. et al.*	
C.4		Origin and Kinematic History of Highly Strained Gneisses in the Eastern Jan Lake Area *Shi, R. & Lewry, J.F.*	(F,M)
C.5		Petrography, Fluid Inclusions and Stable Isotope Data from the Bakos Lode Gold Deposit *Fayek, M. et al.*	
C.6		Mineral Parageneses, Fluid Characteristics and Radiogenic Isotope Data from the Proterozoic Gold Zone, La Ronge Domain *Hrdy, F. et al.*	
C.7		Tourmaline, Phosphate Minerals, Zircon and Pitchblende in the Athabasca Group: Maw Zone and McArthur River areas, Saskatchewan *Quirt, D. et al.*	
C.8		The East Athabasca Mylonite Zone: An Archean Segment of the Snowbird Tectonic Zone *Hanmer, S. et al.*	

---

*	Database Number (formerly project number, used for samples etc.)
F	Field component in 1991
W	Project covered wide area (not shown on map)
C	Contract project in 1991
M	Map(s) or chart(s) in accompanying package
G	Federal (Geological Survey of Canada) component
N	Report not included in this publication

A. Northern Projects Under the Canada-Saskatchewan Partnership Agreement on Mineral Development 1990-95, in Year 1991-92

Revision Bedrock Geological Mapping: Bootleg Lake-Birch Lake Area (Parts of NTS 63K-12 and 63L-9)[1]

David J. Thomas

Thomas, D.J. (1991): Revision bedrock geological mapping: Bootleg Lake-Birch Lake area (parts of NTS 63K-12 and 63L-9); *in* Summary of Investigations 1991, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 91-4.

Geological field work during 1991 completed the third consecutive season of 1:12 500 scale mapping in the greater Flin Flon-Amisk Lake area. New mapping covered the area between Bootleg and Birch Lakes to the southwest of the previously mapped Douglas Lake-Phantom Lake area (Thomas, 1989, 1990). Adjacent 1:12 500 scale geological mapping was undertaken around Kaminis Lake (Reilly, 1990, this volume) and Table Lake (Slimmon, this volume).

Previous mapping by the Geological Survey of Canada throughout the greater Flin Flon-Amisk Lake region includes work by: Tyrrell (1902), McInnes (1909), Bruce (1918), Alcock (1923), Ambrose (1936a, 1936b), Tanton (1941a, 1941b), Stockwell (1960) and Heywood (1966). More recent federal and provincial investigations were undertaken by Byers *et al.* (1965), Pearson (1979a, 1979b, 1981, 1983, 1984) and Galley and Franklin (1987). The regional geology and metallogeny of the Flin Flon area has been reviewed by: Stauffer and Mukherjee (1971), Stauffer *et al.* (1975), Price (1977), Coombe (1979, 1984), Pearson and Galley (1985), Pearson *et al.* (1986), Syme (1986, 1987, 1988), and Bailes and Syme (1987, 1989).

1. Regional Geology

The Bootleg Lake-Birch Lake area lies within the Aphebian volcano-plutonic Flin Flon-Snow Lake greenstone belt, extending for about 300 km from Snow Lake, Manitoba through Flin Flon to the Hanson Lake-Bigstone Lake area in Saskatchewan. The greenstone belt over most of its length is lithologically and metamorphically transitional northwards into the higher grade Kisseynew Gneiss belt (Ashton *et al.*, 1986) and to the south is unconformably overlain by Ordovician rocks of the Winnipeg Sandstone and Red River Dolomite.

The Amisk Group, the oldest group of rocks in the region, comprises a metamorphosed assemblage of predominantly subaqueous mafic flows and flow breccias, subordinate felsic flows and associated felsic porphyries, lesser volcaniclastic and sedimentary rocks, and a suite of syn- to post-volcanic dykes and sills. The group is unconformably overlain by metamorphosed alluvial sandstones and conglomerates of the Missi Group. Post-Missi intrusions include the Boundary intrusions, a suite of ultramafic to felsic composition (Stockwell, 1946, 1960; Syme, 1975; Syme and Forester, 1977) and the younger composite gabbroic to granitic Boot Lake-Phantom Lake Complex (Galley and Franklin, 1987).

Bailes and Syme (1987, 1989) recognize one phase of pre-Missi folding (P1) and four phases of post-Missi folding (P2 to P5) in the Flin Flon-White Lake area. Strong north-, northeast- and northwest-trending fault systems are predominant features, and in some cases are identified as boundaries separating allocthonous blocks of distinctively different stratigraphic and metamorphic character (Bailes and Syme, 1987, 1989; Syme, 1986, 1987, 1988). Thomas (1990) and Reilly (1990) described a major northwesterly-trending structural zone, interpreted as an "early" pre-Missi northeasterly-directed thrust extending from the West Arm of Schist Lake through the Bootleg Lake area. This structure separates volcano-plutonic sequences of markedly contrasting lithological and structural styles.

U-Pb zircon ages obtained from a felsic crystal tuff in the Flin Flon area indicates Amisk volcanism occurred around 1886 Ma (Syme *et al.*, 1987). A similar age of 1888 ±12 Ma (Bickford *et al.*, 1986) has been obtained from a felsic porphyry at Hanson Lake which was previously interpreted as a rhyolite flow. More recent U-Pb zircon work on a sheared tonalite-granodiorite from the Mystic Lake area has yielded an age of 1908 ±3 Ma (Heaman *et al.*, this volume) identifying it as the oldest intrusive yet recognized in this part of the belt and thereby making the surrounding volcanics even older. Possible synvolcanic plutons elsewhere in the Flin Flon-Snow Lake belt yield U-Pb zircon ages as follows: Richard Lake tonalite, 1889 +8/-6 Ma (Bailes *et al.*, 1988), Sneath Lake tonalite, ca. 1886 Ma (Bailes, 1990), Herblet Lake gneiss dome 1890 +8/-6 Ma (Gordon *et al.*, 1990), and a poorly constrained age from the Cliff Lake Pluton, 1871 +32/-25 Ma (Gordon *et al.*, 1990).

Emplacement of post-volcanic plutons in the greater Flin Flon-Amisk Lake area occurred between 25 to 50 Ma later based on U-Pb data for the Annabel Lake granodiorite, (1860 ±6 Ma; Ansdell and Kyser, 1990), the post-Missi intrusions such as the Phantom Lake "granite" (1841 +3/-2 Ma; Heaman *et al.*, this volume) and Boundary Intrusions (1834 +4/-3 Ma; Heaman *et al.*, this volume). Detrital zircons from the Missi Group

---

(1) Saskatchewan project A.111, from which this report derives, was funded in 1991 under the Canada-Saskatchewan Partnership Agreement on Mineral Development 1990-95.

dated by the single zircon Pb-evaporation technique yielded ages between 1869 ±9 to 2092 ±38 Ma although most fall in the interval 1900 to 1880 Ma (Ansdell *et al.*, 1991). Deposition of the Missi Group is therefore constrained between the youngest detrital zircon age of 1869 ±9 Ma and the intrusion of the Phantom Lake "granite" at 1841 +3/-2 Ma.

2. Geology of the Bootleg Lake-Birch Lake Area

The Bootleg Lake-Birch Lake area is underlain by a variably strained volcanic sequence composed predominantly of mafic volcanic flows and minor fragmentals termed the **Mystic Lake Assemblage**. These volcanics are intruded by an "older" suite comprising tonalitic, granodioritic and quartz dioritic rocks collectively termed the **Mystic Lake Tonalite** (1908 ±3 Ma, Heaman *et al.*, this volume). Large sill-like intrusions of quartz diorite to gabbro, and locally of ultramafic composition intrude the volcanic rocks and Mystic Lake Tonalite. The volcanic and plutonic rocks have undergone heterogeneous strain with many of the rocks exhibiting submylonitic to mylonitic or phyllonitic fabrics.

The Mystic Lake Assemblage is intruded by the Reynard Lake Pluton, a large multiphase to composite intrusion, which post-dates the strong penetrative S_1 fabric developed throughout the volcanic and earlier plutonic rocks. The Reynard Lake Pluton has been variably foliated by later tectonic events. An age of 1853 ±8 Ma has been obtained by single zircon Pb-evaporation technique from a late phase of the Reynard Lake Pluton (Ansdell and Kyser 1990).

a) Metavolcanic Rocks

Volcanic and plutonic rocks of the Mystic Lake Assemblage occur in two main belts, east and west of the Reynard Lake Pluton. The northwesterly-trending **eastern belt** of mafic flows and gabbros, ranges from 300 to 1600 m wide, and extends from east of Wekach Lake to Bootleg Lake and Meridian Creek, and is largely continuous with a similar sequence of rocks in the Mystic Lake area (Figure 1). The two areas are connected by a narrow wedge of volcanic rocks constricted between the Reynard Lake Pluton and the Boot Lake-Phantom Lake Intrusion along Meridian Creek. Immediately east of the south end of Wekach Lake, the Mystic Lake Assemblage is intruded by gabbroic and dioritic rocks of the Boot Lake-Phantom Lake Intrusion (Thomas, 1989). Mafic volcanics and large gabbro-diorite sills between the Kaminis Lake Pluton and the West Arm of Schist Lake are interpreted to be the southeastern continuation of the assemblage.

Mineral assemblages in volcanic rocks of the eastern belt indicate lower amphibolite facies metamorphic grades were attained. To the east of the Reynard Lake Pluton, notably northeast of Wekach Lake, highly-strained mafic flows and gabbroic sills of the Mystic Lake Assemblage are in tectonic contact and structurally overlie lesser deformed and metamorphosed (greenschist facies) volcaniclastic rocks and minor flows of the Douglas Lake Assemblage (Thomas, 1989).

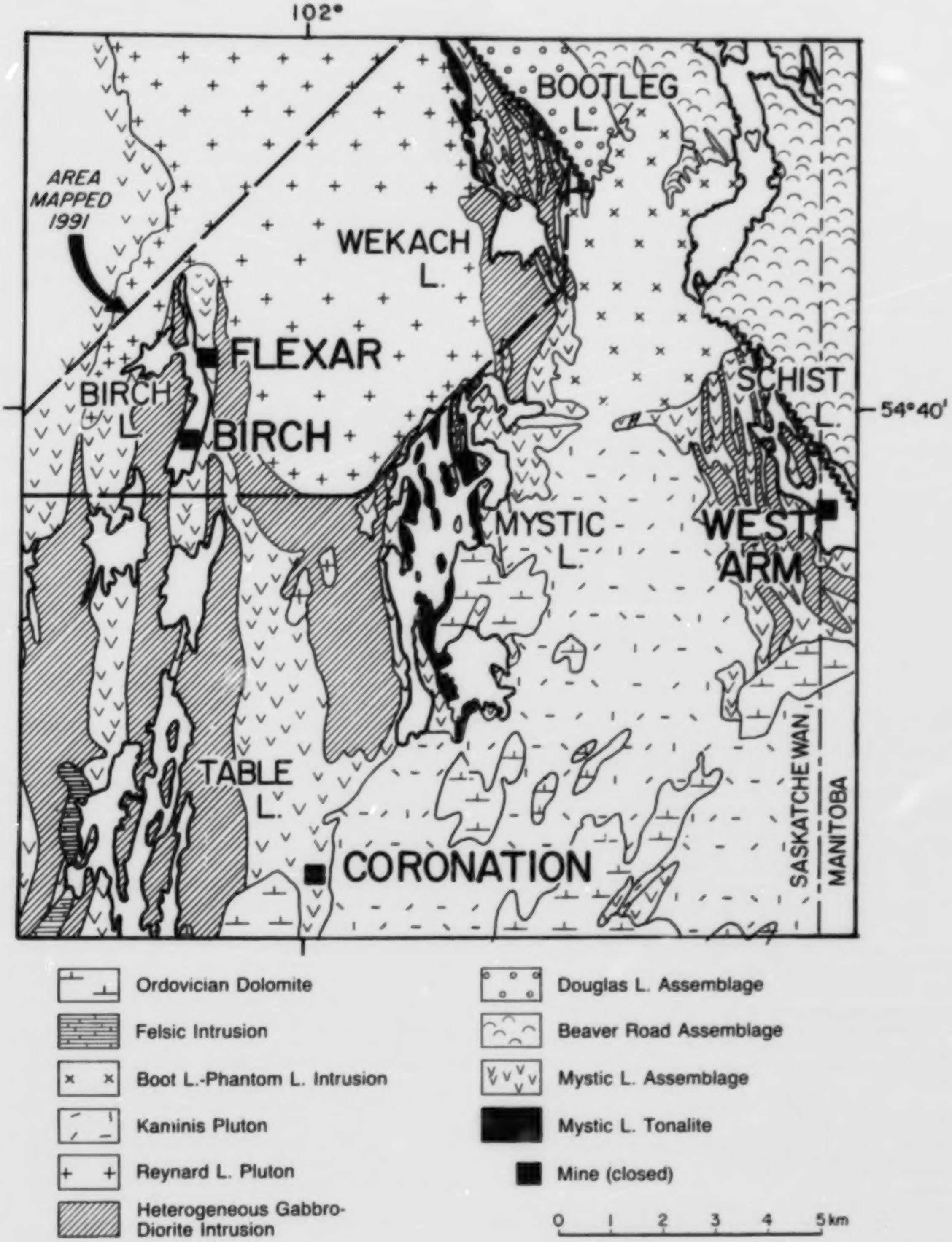
Gross compositional layering ranges from subvertical to steeply southwest dipping throughout the eastern belt. Primary volcanic textural and structural features are rarely preserved; however, the limited way-up data available suggests a southwest facing homoclinal sequence at least 1200 m thick. Within the belt deformation increases southwestwards from Bootleg Lake to Wekach Lake and northwards along Meridian Creek. Rocks along Becker Bay at the north end of Wekach Lake are submylonitic to mylonitic or phyllonitic, in marked contrast, to weakly cleaved and less metamorphosed rocks northeast of Meridian Creek and Bootleg Lake.

The **western belt**, comprising mafic volcanics and mafic to ultramafic intrusions, is a south-trending sequence extending from Birch Lake through Table Lake and eventually below the Phanerozoic cover (Slimmon, this volume; Reilly, this volume). Only the most northerly portion of this belt, where metamorphic grade increases from greenschist facies in the west to amphibolite facies in the east, is described in this report (i.e. Birch-Echo-Konuto Lakes).

Primary volcanic textural and structural features are poorly preserved within the strongly deformed mafic volcanic sequence between Birch Lake and the western margin of the Reynard Lake Pluton. Pillow-like shapes and quartz-feldspar amygdales are locally observed. A distinctive heterolithologic mafic tuff-breccia to lapilli tuff unit outcrops semicontinuously along the central part of Birch Lake.

East of Birch Lake, mafic to intermediate rocks of either flow or tuffaceous origin are strongly deformed (mylonitic to phyllonitic) and typically display a streaky tectonic banding defined by variably silicified and epidotized domains within the mylonitic metavolcanics. Along the east and west shores of Birch Lake, and in part on the large island at the south end of the lake, broadly conformable, discontinuous zones of intense silicification are developed within mafic flows. These silicified zones may correlate with the same stratigraphic interval which hosts massive sulphide mineralization at the Birch and Flexar mines.

West of Birch Lake, a 1000 m wide zone of mafic metavolcanics occurs between two large mafic sill-like intrusions. These rocks can be traced for 15 km from Birch Lake through Konuto Lake to Table Lake. They are typically aphanitic to fine grained, dark grey to greenish black, massive, and exhibit subconchoidal fractures. Primary volcanic structures are generally not discernible, although the rocks are variably amygdaloidal (0 to 20 percent) and locally feldspar phyric. Net-veining by epidote, plagioclase, quartz and locally carbonate is pervasive. Contact metamorphism and alteration related to the large mafic sills may provide an explanation of the cherty "hornfelsic" and veined characteristics of these rocks.



102°
AREA MAPPED 1991
BOOTLEG L.
WEKACH L.
FLEXAR
BIRCH L.
BIRCH
SCHIST L.
54°40'
MYSTIC L.
WEST ARM
TABLE L.
CORONATION
SASKATCHEWAN
MANITOBA
Ordovician Dolomite
Felsic Intrusion
Boot L.-Phantom L. Intrusion
Kaminis Pluton
Reynard L. Pluton
Heterogeneous Gabbro-Diorite Intrusion
Douglas L. Assemblage
Beaver Road Assemblage
Mystic L. Assemblage
Mystic L. Tonalite
Mine (closed)
0 1 2 3 4 5 km

Figure 1 - Generalized geology of the Flin Flon-East Amisk Lake area.

b) Mystic Lake Tonalite

The Mystic Lake Tonalite, which is strongly deformed, comprises granodiorite, tonalite and quartz diorite intruding finely layered mafic-felsic metavolcanic rocks at the north end of Wekach Lake. This intrusive suite, identical to that described in the Mystic Lake area by Reilly (1990), outcrops in a narrow zone that extends at least 1.5 km northwest of Wekach Lake and to the west is intruded by the Reynard Lake Pluton. The eastern margin of the Reynard Lake Pluton locally contains abundant xenoliths of highly strained (mylonitic) Mystic Lake Tonalite. Felsic layers, within the surrounding layered mafic-felsic volcanics, become thicker and more prevalent towards the Mystic Lake Tonalite, thereby suggesting they represent a dyke injection aureole around the intrusion. Later tectonism has rotated these dykes subparallel to foliation resulting in a marked compositional layering, which previous workers interpreted as primary layering of mafic and felsic tuffs (Byers *et al.*, 1965; Pearson, 1979a).

c) Heterogeneous Gabbro-Diorite Intrusions

Metavolcanic rocks in the Wekach Lake and Birch Lake areas are intruded by numerous large heterogeneous sills ranging in composition from diorite through gabbro to pyroxenite. Individual intrusions are compositionally and texturally variable with grain sizes ranging from fine to coarse, and locally pegmatitic over a few metres. Brecciation and stoping of earlier phases by successively younger phases, as well as rhythmic mineral layering and cumulate-like textures are common in many of the larger bodies. Contacts between the various compositional and textural varieties are both gradational over several centimetres, as well as sharp and crosscutting. The contact between intrusions and the metavolcanics is generally sharp, although, transitional zones (10 to 200 m wide) of intercalated volcanics and narrow mafic sills typify the margins of many bodies. Similarly, broad bands of metavolcanic rocks are common throughout the intrusions. Individual intrusions range in width from several tens of metres to greater than 1 km and can be traced along strike for more than 10 km. The layered and multiphase nature of the intrusions and their intercalated relationship with the volcanics are characteristics of sheeted dyke/sill complexes found in oceanic or back-arc rift environments.

d) Reynard Lake Pluton

The Reynard Lake Pluton is a large tear-drop shaped intrusion approximately 14 km in length and 4.5 km wide at its southern end. Only the southern third of the pluton was mapped during 1991 (Figure 2). The pluton is multiphase, and possibly in part composite, ranging from early diorite to progressively more felsic younger

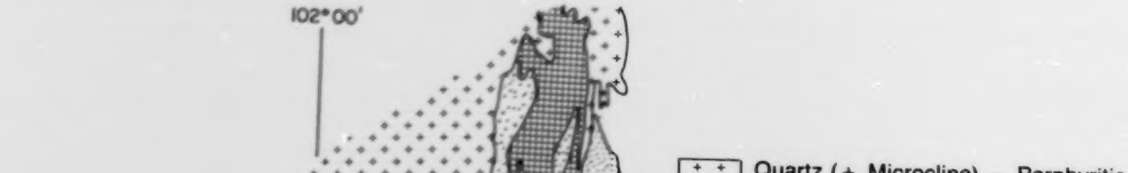


Figure 2 - Simplified geology of the southern portion of the Reynard Lake Pluton situated between Wekach Lake and Bootleg Lake.

phases. It is unclear if all intrusive phases are genetically related and part of a single fractionated series.

An early, marginal phase is discontinuously developed along the western, and in part eastern, margins of the pluton. It comprises a heterogeneous assemblage of fine- to medium-grained melanocratic to leucocratic diorite, quartz diorite, monzodiorite and locally gabbro. A massive, medium-grained, light greenish-grey feldspar diorite porphyry forms a distinct member of this marginal phase in the Birch Lake area. Xenoliths of metavolcanic rocks are prevalent in the marginal phase which is itself locally intruded by up to 30 percent younger phases of the pluton.

The main phase of the pluton is also distinctive being composed of fine- to medium-grained, equigranular, leucocratic rocks ranging from diorite through quartz diorite to tonalite/granodiorite. Hornblende varies from 30 percent in the diorite to 10 percent in the tonalite/granodiorite members; biotite ranges from 0 to 5 percent. Although, the various rocks comprising the main phase exhibit a compositional continuum, their contacts are generally sharp, irregular and intrusive on a small scale, but at a mappable level are transitional, marked by broad zones (2 to 50 m wide) of pronounced dykes and blocky agmatitic blocky intrusion breccias. The main phase intrusives show a weak foliation marked by a poor alignment of hornblende and biotite.

The main phase of the pluton is intruded by several distinctive leucocratic granodioritic phases which are presumed to be of broadly similar age. Along the southern and eastern margin of the pluton, a narrow arcuate body of fine- to medium-grained, light grey to yellowish-grey, weakly to moderately foliated and rodded, biotite (<5 percent) granodiorite intrudes fine-grained diorites and quartz diorites. The eastern margin of the pluton is intruded by several irregular to sheet-like bodies of greyish-pink to greyish-orange pink, medium-grained, moderately foliated biotite (±hornblende) granodiorite. Mafic mineral content ranges from 5 to 10 percent. Broad dyke injection aureoles (>30 percent dykes and veins) up to several hundred metres wide, envelop and are transitional into the main bodies of this rock.

The youngest major phase of the Reynard Lake Pluton is a medium- to coarse-grained, greyish-orange to pale orange quartz- (±microcline)-porphyritic granodiorite to quartz monzonite. It outcrops in the north-central part of the pluton and crosscuts internal contacts between older phases. Quartz-porphyritic granodiorite dykes are abundant peripherally around the body. Along its western side they typically form westerly-trending radial swarms, whereas on the east side they occur in northerly-trending conformable to concentric swarms. The body exhibits a gradational zonation from an outer zone of quartz-porphyritic rocks through quartz- and sparsely microcline-porphyritic rocks to a coarsely microcline + quartz-porphyritic core. Within the microcline-porphyritic core, phenocrysts compose 5 to 10 percent of the rock and range from 0.5 to 4.0 cm long. Pale blue opalescent quartz phenocrysts, ranging from 0.5 to 1.5 cm diameter, compose up to 15 percent of the rock. Biotite (2 to 7 percent) is the principal mafic mineral, although hornblende (2 to 5 percent) is locally dominant.

Foliation is generally weakly developed and quartz phenocrysts are moderately-elongated (rodded) in the core of the body. Subrounded xenoliths of quartz diorite and granodiorite commonly containing quartz megacrysts, are particularly abundant in the western margin. These megacrysts display a progressive decrease in size and abundance towards the core of the individual xenoliths.

The youngest rocks identified are greyish-pink to moderate-orange pink, fine-grained aplitic dykes which are generally confined to the north central and western parts of the pluton.

3. Economic Geology

Mineral exploration and development for both base and precious metals has been on going in the Flin Flon-Amisk Lake area since the early part of this century. The area hosts numerous gold occurrences of which several have realized a small amount of production (Newcor, Bootleg, Henning-Maloney, Monarch, and Phantom). Most of the gold occurrences have been geologically mapped and described and the gold metallogeny of the region reviewed by Pearson (1979b, 1980, 1981, 1983, 1984), Pearson and Galley (1985), Pearson *et al.* (1986), Coombe (1984), and Galley and Franklin (1987). The majority of gold occurrences can be classified as shear-hosted sulphide-bearing auriferous quartz veins. A number of Au-Cu (±W±Mo) occurrences in the Phantom Lake area exhibit characteristics broadly suggestive of porphyry-type mineralization related to a calc-alkaline felsic intrusion (Phantom Lake "granite"). A possible example of an Early Proterozoic epithermal precious metal deposit occurs on Missi Island in Amisk Lake (Laurel Lake deposit) where fracture-controlled gold-silver vein-type mineralization is associated with a rhyolite porphyry and felsic fragmentals.

Within the map area, five gold occurrences are listed and described in the Saskatchewan Mineral Index File system. They include the Angelski, Wekach Lake Gold Mines, Dodo, Flash and VM Group showings. Evidence of unrecorded prospecting, sampling and trenching was encountered throughout the area, most notably between Reynard and Birch lakes.

Two past base metal producers, the Birch and Flexar mines, are also situated in the area; a third past

Table 1 - Production History of Base Metal Producing Mines in the Birch Lake Area. Production figures obtained from Coombe (1991).

Mine	Period	Tons	Grade	
			Cu%	Zn%
Birch	1957-60	307 265	6.15	—
Flexar	1969-72	337 146	3.75	0.47
Coronation	1960-65	1 412 861	4.25	0.30

producer, the Coronation mine, is located about 9 km south of the Birch Lake Mine (Table 1). These relatively small tonnage, Cu-rich Zn-poor massive sulphide deposits have been described by Byers (1969) and more recently by Coombe (1991).

4. Selected Bibliography

Alcock, F.J. (1923): Flin Flon map area, Manitoba and Saskatchewan; Geol. Surv. Can., Summ. Rep., pt C, p1-36.

Ambrose, J.W. (1936a): Progressive kinetic metamorphism in the Missi Series, near Flin Flon, Manitoba; Am. J. Sci. (5th ser.), v32, p257-286.

__________ (1936b): Structures in the Missi Series near Flin Flon, Manitoba; Trans. R. Soc. Can. (3rd ser.), v30, p81-98.

Ansdell, K.M. and Kyser, T.K. (1990): Age of granitoids from the western Flin Flon Domain: an application of the single-zircon Pb-evaporation technique; *in* Summary of Investigations 1990, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 90-4, p136-142.

Ansdell, K. M., Stauffer, M.R., Kyser T.K. and Edwards, G.(1991): Age and source of detrital zircons from the Missi Group: a Proterozoic molasse deposit, Trans-Hudson Orogen (abstract); GAC-MAC Joint Annual Meeting, May 1991, Toronto, Prog. Abstr.

Ashton, K.E., Wheatley, K.J., Moser, D., Paul, D. and Wilcox, K.H. (1986): The Kisseynew Gneisses of Saskatchewan: update; *in* Summary of Investigations 1986, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 86-4, p145-148.

Bailes, A.H. (1990): Chisel-Anderson Project (NTS 63K/16SE); *in* Man. Energy Mines, Report of Field Activities 1990, p.43-48.

Bailes, A. H. and Syme, E.C. (1987): Geology of the Flin Flon-White Lake area; GR87-1-1, Man. Energy Mines.

__________ (1989): Geology of the Flin Flon-White Lake area, Man. Energy Mines, Geol. Rep. GR87-1, 313p.

Bailes, A. H., Gordon, T.M. and Hunt, P.A. (1988): U-Pb zircon geochronology of the Richard Lake Tonalite, a possible synvolcanic pluton in the Snow Lake area; *in* Man. Energy Mines, Report of Field Activities 1988, p63-65.

Bickford, M.E., Van Schmus, W.R., Macdonald, R., Lewry, J.F., and Pearson, J.G. (1986): U-Pb zircon geochronology project for the Trans-Hudson: current sampling and recent results; *in* Summary of Investigations 1986, Saskatchewan Geological Survey; Sask. Energy Mines, Misc. Rep. 86-4, p101-107.

Brownell, G.M. and Kinkel, A.R. (1935): The Flin Flon Mine, geology and paragenesis of the ore deposit; Trans. Can. Inst. Min. and Metall., v38, p261-286.

Bruce, E.L. (1918): Amisk-Athapapuskow Lake District; Geol. Surv. Can., Mem. 105, 91p.

Byers, A.R., ed. (1969): Symoposium on the Geology of the Coronation Mine, Saskatchewan; Geol. Surv. Can., Pap. 68-5, 329p.

Byers, A.R., Kirkland, S.J.T. and Pearson, W.J. (1965): Geology and mineral deposits of the Flin Flon area, Saskatchewan; Sask. Dep. Miner. Resour., Rep. 62, 95p.

Coombe, W. (1979): Mineral deposits and regional metallogeny, southeastern Shield; *in* Summary of Investigations 1979, Saskatchewan Geological Survey; Sask. Miner. Resour., Misc. Rep. 79-10, p120-133.

__________ (1984): Gold in Saskatchewan; Sask. Energy Mines, Open File Rep. 84-1, 134p.

__________ (1991): Base Metals in Saskatchewan; Sask. Energy Mines, Open File Rep. 91-1, 218p.

Gale, G.H. (1981): Mineral Deposit studies in the Flin Flon area; Man. Miner. Resour. Div., Report of Field Activities 1981, p45-55.

Gale, G.H. and Eccles, D.R. (1988): Mineral deposits and occurrences in the Flin Flon area NTS 63K/13SW: Part 2, Flin Flon area (63K/13SW); Man. Energy Mines Miner. Deposit Ser., Rep. 2, 98p.

Galley, A.G. and Franklin, J.M. (1987): Geological setting of gold, copper, tungsten and molybdenum occurrences in the Phantom Lake region; *in* Summary of Investigations 1987, Saskatchewan Geological Survey; Sask. Energy Mines, Misc. Rep. 87-4, p115-124.

__________ (1989): Setting of gold in the Phantom Lake region, Saskatchewan; *in* Galley, A.G. (ed.), Investigations by the Geological Survey of Canada in Manitoba and Saskatchewan during the 1984-1989 Mineral Development Agreements; Geol. Surv. Can., Misc. Rep. 2133, p33-44.

Gordon, T.M., Hunt, P.A., Loveridge, W.D., Bailes, A.H. and Syme, E.C. (1987): U-Pb zircon ages from the Flin Flon and Kisseynew belts, Manitoba: Chronology of Early Proterozoic crust formation; GAC-MAC Joint Annual Meeting, May 1987, Saskatoon, Prog. Abstr., v12, p48.

Gordon, T.M., Hunt, P.A., Bailes, A.H. and Syme, E.C. (1990): U-Pb zircon ages from the Flin Flon and Kisseynew belts, Manitoba: chronology of crust formation at an Early Proterozoic accretionary margin; *in* Lewry, J.F. and Stauffer, M.R. (eds.), The Early Proterozoic Trans-Hudson Orogen of North America: Geol. Assoc. Can., Spec. Pap. 37, p177-199.

Heywood, W.W. (1966): Precambrian geology of the Ledge Lake area, Manitoba and Saskatchewan, Canada; unpubl. Ph.D. thesis, Univ. Wash.

Koo, J. and Mossman, D.J. (1975): Origin and metamorphism of the Flin Flon stratabound Cu-Zn sulphide deposit, Saskatchewan and Manitoba, Canada; Econ. Geol., v70, p48-62.

MacQuarrie, R.R. (1980): Absolute age of the Flin Flon deposit; CIM Bull., abstr., v73, no. 820, p53.

McInnes, W. (1909): Explorations on the Churchill River and South Indian Lake; Geol. Surv. Can., Summ. Rep. 1908, p87-92.

Parslow, G.R. and Gaskarth, J.W. (1988): Proterozoic rocks of east-central Saskatchewan: geochemistry, structure and mineralization controls; *in* Summary of Investigations 1988, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 88-4, p127-139.

Pearson, J.G. (1979a): Flin Flon gold prospects: Douglas Lake-Wekach Lake area; 1:10 000 scale prelim. map *with* Summary of Investigations 1979, Saskatchewan Geological Survey, Sask. Miner. Resour., Misc. Rep. 79-10.

__________ (1979b): Flin Flon gold prospects; *in* Summary of Investigations 1979, Saskatchewan Geological Survey, Sask. Miner. Resour., Misc. Rep. 79-10, p145-151.

__________ (1980): Flin Flon gold project; *in* Summary of Investigations 1980, Saskatchewan Geological Survey, Sask. Miner. Resour., Misc. Rep. 80-4, p70-80.

__________ (1981): Gold metallogenic studies: Phantom Lake-Schist Lake area; *in* Summary of Investigations 1981, Saskatchewan Geological Survey, Sask. Miner. Resour., Misc. Rep. 81-4, p97-100.

__________ (1983): Gold metallogenic studies: Flin Flon-Amisk Lake area; *in* Summary of Investigations 1983, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 83-4, p67-74.

__________ (1984): Gold metallogenic studies, the Rio deposit; *in* Summary of Investigations 1984, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 84-4, p123-126.

Pearson, J.G. and Galley, A.G. (1985): Gold deposits of the Flin Flon-Amisk Lake area; *in* Summary of Investigations 1985, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 85-4, p83-89.

Pearson, J.G., MacDougall, F.H., and Galley, A.G. (1986): Geology and evolution of gold occurrences in the Flin Flon-Amisk Lake area, Saskatchewan; *in* Clark, L.A. (ed.), Gold in the Western Shield; CIM, Spec. Vol. 38, p399-411.

Price, D.P. (1977): Geology and economic potential of the Flin Flon-Snow Lake areas; Univ. Man., Cent. Prec. Stud., Annu. Rep., p52-83.

Reilly, B.A. (1990): Bedrock geological Mapping, Mystic-West Arm, Schist Lakes area (northwest part of NTS 63K-12); *in* Summary of Investigations 1990, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 90-4, p25-35.

Sangster, D.F. (1972): Isotopic studies of ore-leads in the Hanson Lake-Flin Flon-Snow Lake mineral belt, Saskatchewan and Manitoba; Can. J. Earth Sci., v9, p500-513.

Stauffer, M.R. and Mukherjee, A. (1971): Superimposed deformations in the Missi meta-sedimentary rocks near Flin Flon, Manitoba; Can. J. Earth Sci., v8, p217-242.

Stauffer, M.R., Mukherjee, A.C. and Koo, J. (1975): The Amisk Group: an Aphebian(?) island arc deposit; Can. J. Earth Sci., v12, p2021-2035.

Stockwell, C.H. (1946): Flin Flon-Mandy area, Manitoba and Saskatchewan; Geol. Surv. Can., Pap. 46-14, 5p.

__________ (1960): Flin Flon-Mandy, Manitoba and Saskatchewan; Geol. Surv. Can., Map 1078A.

Syme, E.C. (1975): Petrogenesis of the boundary intrusions, Flin Flon area, Saskatchewan-Manitoba; unpubl. M.Sc. thesis, Univ. Sask.

__________ (1986): Schist Lake area, Athapapuskow project; *in* Report of Field Activities 1986, Man. Energy Mines, Misc. Rep., p36-42.

__________ (1987): Athapapuskow Lake Project; *in* Report of Field Activities 1987, Man. Energy Mines, Misc. Rep., p30-39.

__________ (1988): Athapapuskow Lake Project; *in* Report of Field Activities 1988, Man. Energy Mines, Misc. Rep. p20-34.

Syme, E.C., Bailes, A.H., Gordon, T.M. and Hunt, P.A. (1987): U-Pb zircon geochronology in the Flin Flon Belt: age of Amisk volcanism; *in* Report of Field Activities 1987, Man. Energy Mines, Misc. Rep; ptGS-19, p105-107.

Syme, E.C. and Forester, R.W. (1977): Petrogenesis of the Boundary Intrusions in the Flin Flon area of Saskatchewan and Manitoba; Can. J. Earth Sci., v14, p444-455.

Tanton, T.L. (1941a): Flin Flon, Saskatchewan and Manitoba; Geol. Surv. Can., Map 632A.

__________ (1941b): Schist Lake, Saskatchewan-Manitoba; Geol. Surv. Can., Map 633A.

Thomas D.J. (1989): Geology of the Douglas Lake-Phantom Lake area (part of NTS 63K-12 and -13); *in* Summary of Investigations 1989, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 89-4, p44-54.

__________ (1990): New perspective on the Amisk Group and regional metallogeny, Douglas Lake-Phantom Lake area, northern Saskatchewan; *in* Summary of Investigations 1990, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 90-4, p. 13-20.

Tyrrell, J.B. (1902): Explorations in the northeastern portion of the district of Saskatchewan and adjacent parts of the district of Keewatin; Annual Report; Geol. Surv. Can., v8, ptF, 48p.

Watters, B.R. (1989): Lithogeochemical studies in the Flin Flon-Amisk Lake area and the Glennie Domain; *in* Summary of Investigations 1989, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 89-4 p55-57.

Revision Bedrock Geological Mapping: Mystic Lake-Kaminis Lake Area (Parts of NTS 63K-12 and 63L-9)[1]

B.A. Reilly

Reilly, B.A. (1991): Revision bedrock geological mapping: Mystic Lake-Kaminis Lake area (Parts of NTS 63K-12 and 63L-9); *in* Summary of Investigations 1991, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 91-4.

This year's bedrock geological mapping in the Mystic Lake-Kaminis Lake area marks the completion of field work initiated by the author in 1990 (Reilly, 1990) in the Mystic Lake area. The Mystic Lake project is part of the larger multi-disciplinary, multi-year Project Seagull and includes bedrock geological mapping by colleagues D. Thomas (1989, 1990, this volume) and B. Slimmon (this volume).

Approximately 55 km^2 were mapped in detail (1:12 500 scale) in the Mystic Lake-West Arm, Schist Lake area during the 1990 field season (Reilly, 1990). Mapping during the summer of 1991 at the same scale included the area west of Mystic Lake to Table Lake, the area around Probert, Kaminis and West Arm, Athapapuskow Lakes along the shield edge to the south and the area north of Mystic Lake to Boot Lake. Approximately 80 km^2 were mapped.

A major objective is to better understand the controls on base metal and gold deposits in the region. Emphasis is placed on the application of physical volcanology and lithogeochemistry in refining the volcanic stratigraphy, mapping and geochemical characterization of intrusive rocks, as well as detailed structural analysis of the complex fault/shear systems transecting the region. Relative ages for volcanism, plutonism and deformation events based on field relationships are being tested by high precision geochronological methods.

1. Previous Work

Previous work in the area is summarized below. This list is far from exhaustive and concentrates on geological survey activity rather than the many topical studies in the area. Tyrrell (1902), McInnes (1909) and Bruce (1916, 1918a, b) recorded the first reconnaissance geological surveys in the Flin Flon-Amisk Lake region. These surveys were followed by detailed mapping by Alcock (1923), Ambrose (1936a, b), Tanton (1941a, b), Stockwell (1946, 1960), Byers and Dahlstrom (1954), Byers *et al.* (1965), Heywood (1966) and Froese (1968). Recent 1:12 500 scale mapping by the Saskatchewan Geological Survey in the area includes summaries by Thomas (1989, 1990, this volume), Reilly (1990) and Slimmon (this volume). The Manitoba Geological Survey have just completed 1:15 840 scale mapping along the Saskatchewan/Manitoba border and results can be found in a report by Bailes and Syme (1987) and summaries by Syme (1985, 1986, 1987, 1988).

2. Regional Geology

The area around Mystic Lake forms part of the Flin Flon-Amisk Lake region of the Flin Flon-Snow Lake Domain, the most southern exposure of the Trans-Hudson Orogen (Lewry and Collerson, 1990; Hoffman, 1990). The Flin Flon greenstone belt comprises an assemblage of predominantly subaqueous and subordinate subaerial Aphebian metavolcanic rocks, associated metasedimentary rocks and dykes/sills designated as the Amisk Group. These, at 1886 Ma (Syme *et al.*, 1987), are the oldest rocks hitherto recognized in the belt. The Amisk Group is unconformably overlain by a sequence of molasse-type metasedimentary rocks of the Missi Group dated at 1830 Ma (Gordon *et al.*, 1987), but these do not crop out in the map area. The supracrustal rocks in the region have been intruded by various intrusions dated at between 1886 Ma and 1820 Ma (MacQuarrie, 1980; Gordon *et al.*, 1987; Ansdell and Kyser, 1991). Major intrusions in the area around Mystic Lake are the Reynard Lake Pluton, Kaminis Lake Intrusion and Mystic Lake Intrusion. Flat-lying Ordovician dolomites unconformably overly the Precambrian rocks to the south.

3. Summary of Major Geological Features

The following major geological features have been recognized in the Mystic Lake area since completion of last year's report:

1. The Mystic Lake Shear Zone (MLSZ) represents early west-side-up thrusting which forms part of a regional northeast-verging imbrication system.
2. A new U-Pb age of 1908 ±3 Ma (Heaman *et al.*, this volume) from the Mystic Lake area represents the oldest Proterozoic rocks found so far in the Flin Flon region.
3. East-younging and structurally dismembered ultramafic and mafic rocks in the area east of Table Lake may represent an ophiolitic sequence.

---

(1) Saskatchewan Project A.112, from which this report derives, was funded in 1991 under the Canada-Saskatchewan Partnership Agreement on Mineral Development 1990-95.

4. A predominantly mafic pyroclastic volcanic rock sequence with subordinate felsic pyroclastic volcanic rocks is found between Barney Lake and Tanyss Lake.

5. The Reynard Lake Pluton has been subdivided into a multiphase, and in part composite granitoid intrusion, and may represent a diapiric intrusive body.

a) Mystic Lake Shear Zone

A broad zone of pervasive ductile shearing, which affects volcanic rocks and the Mystic Lake Intrusion, extends through Mystic Lake with a maximum width of approximately 2.5 km (Reilly, 1990). This zone is termed the Mystic Lake Shear Zone (MLSZ) and represents the earliest recognizable structure in the map area. The MLSZ can be traced through Wekach Lake and as far north as Meridian Lake (Thomas, this volume). Sheared volcanic rocks along the north shore of the West Arm of Athapapuskow Lake suggest that the MLDZ also extends farther south (Figure 1).

Shearing in this zone is defined by a strong to intense planar fabric (S1) which strikes predominantly north-south and dips steeply to the west. The trend of this fabric bends northwesterly near Meridian Lake and easterly near the west arm of Athapapuskow Lake. The foliation is enhanced by millimetre-scale tectono-metamorphic layering, boudinaged and sheared quartz veins, dykes/sills and transposed layering. Grain size gradations due to strain gradients also enhance the shear foliation. Mylonitic textures are exhibited by porphyroclasts, augens and ribbons in all rock types affected by the shearing.

Minor mesoscopic structures associated with the ductile shearing are: 1) a moderate south- to southwest-plunging stretching lineation (L1); 2) rootless isoclinal folds (F1a) which developed early during the shear event and plunge at various angles upon rotation into the stretching lineation; 3) asymmetric shear folds (F1b) which plunge at a high angle to the stretching lineation, and were produced during the waning stages of the shear event; and 4) asymmetric pull-apart structures. The asymmetry of these structures indicate west-side-up, oblique thrusting.

b) Mystic Lake Intrusion

The Mystic Lake Intrusion is predominantly granodiorite to tonalite (Reilly, 1990) and extends at least 12 km from north of Wekach Lake (Thomas, this volume) to the south end of Spectral Lake. It has a maximum width at Mystic Lake of about one kilometre and becomes thinner to the north and south. The intrusion has been deformed apparently in a similar manner to the supracrustal rocks in the Mystic Lake Shear Zone and exhibits mylonitic textures such as quartz augens and ribbons, feldspar porphyroclasts and local tectonic layering due to strain gradients.

A U-Pb zircon age date of 1908 ±3 Ma has been obtained from the intrusion by Heaman *et al.* (this volume). The presence of mafic rafts in the intrusion and the

ORDOVICIAN

⊥ ⊥	Dolomite

PRECAMBRIAN

INTRUSIONS

\ / \ /	Boot Lake-Phantom Lake Intrusion
X X X	Kaminis Lake Intrusion
✢ ✢	Reynard Lake Intrusion (granodiorite)
† † †	Reynard Lake Intrusion (diorite/gabbro)
+ + +	Mystic Lake Intrusion

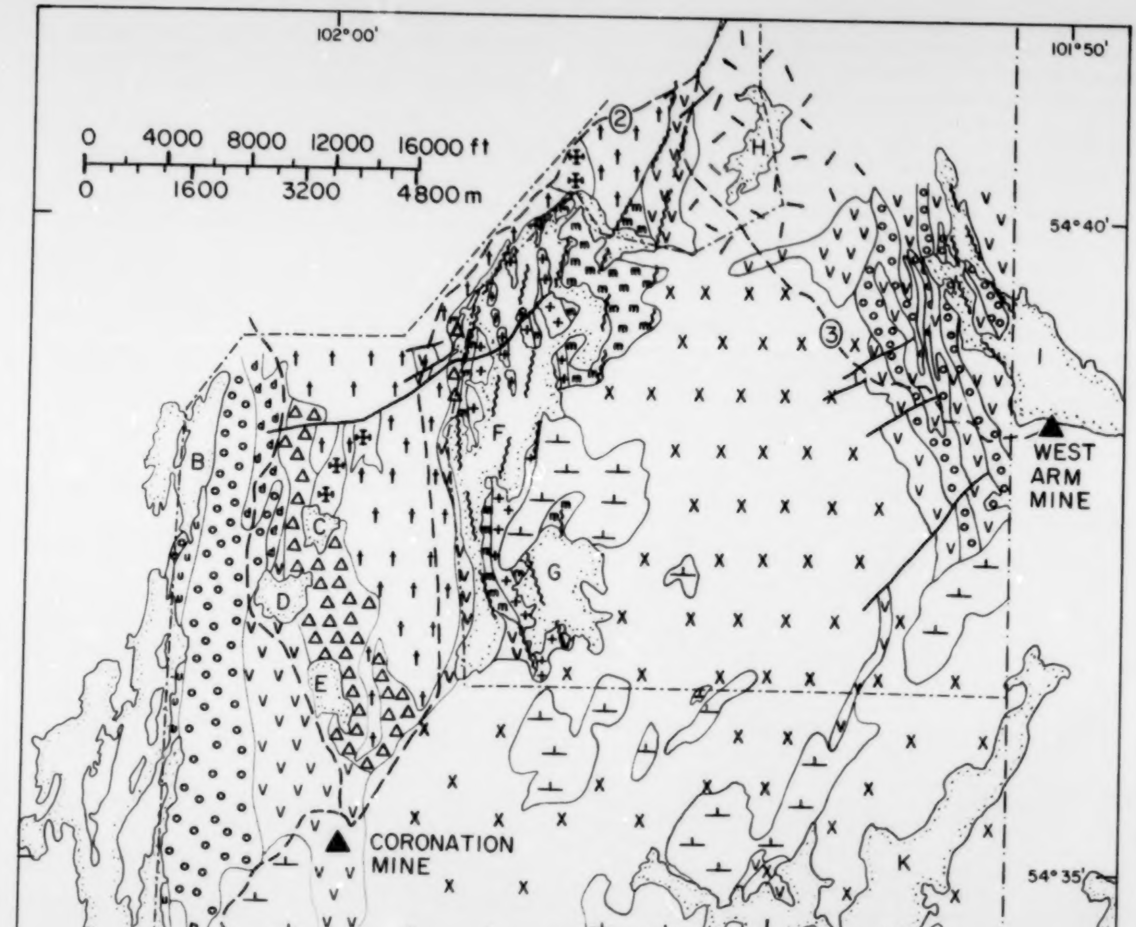
AMISK GROUP

Δ Δ	Mafic Pyroclastic Volcanics (minor felsic volcanics)
V V V	Mafic Flow Volcanics
d d d	Mafic Dyke/Sill Complex
o o o	Dioritic/Gabbroic Sills
u u u	Ultramafic Sills
m m m	Tectonically Layered Volcanics

STRUCTURE

——	Fault
〰	Shear Zone
-·-·-	Limit of Mapping 1991

Figure 1 - Geological sketch map of the Mystic Lake-Kaminis Lake area. A, Table Lake; B, Ruth Lake; C, Barney Lake; D, Stitt Lake; E, Tanyss Lake; F, Mystic Lake; G, Spectral Lake; H, Boot Lake; I, West Arm-Schist Lake; J, Probert Lake; K, Kaminis Lake; L, West Arm-Athapapuskow Lake; M, Leaf Lake; N, Taylor Lake; O, Mariache Lake; (1) Maraiche Lake Road; (2) Mystic Lake Road; (3) West Arm Mine Road.


102° 00'
101° 50'
54° 40'
54° 35'
0 4000 8000 12000 16000 ft
0 1600 3200 4800 m
A
B
C
D
E
F
G
H
I
J
K
L
M
N
O
①
②
③
CORONATION MINE
WEST ARM MINE
SASK.
MAN.

presence of sills/dykes of similar composition as the Mystic Lake Intrusion in the supracrustal assemblage suggests that the Mystic Lake Intrusion intruded the volcanic rocks, and discounts the idea that the intrusion represents underlying basement rock. This implies that the volcanic rocks are older than 1908 Ma. The timing of early thrusting associated with the Mystic Lake Shear Zone can also be bracketed between 1908 Ma and 1853 Ma as the Reynard Lake Pluton crosscuts this structure.

c) Ultramafic-mafic Rock Sequence

An ultramafic-mafic rock sequence lies in the area between Ruth Lake-Table Lake and the Maraiche Lake Road (Figure 1). The sequence strikes north-south, dips steeply to the west and youngs to the east. At the base of the sequence is a series of pyroxenite/peridotite sills which exhibit grain size gradation from fine to coarse. A detailed description of the ultramafic sills is given by Slimmon (this volume). Stratigraphically overlying the ultramafic rocks is a large diorite/gabbro sill approximately 1 km thick. Grain size is predominantly medium, ranging from fine to coarse. Quartz diorite forms part of this mafic intrusion east of Ruth Lake. Overlying the mafic intrusive rocks is a sill/dyke complex which is best exposed between Ruth Lake and Barney Lake along the Maraiche Lake Road. This complex is defined by alternating fine- to medium-grained diorite/gabbro sills/dykes and very fine-grained mafic volcanics/diabase sills/dykes. Both rock types are found in approximately equal proportions with widths typically less than one metre, but ranging from 5 cm up to 3 m. The sill/dyke complex is not continuous in strike length to the south where the diorite/gabbro intrusion is apparently overlain by mafic volcanics. Mafic volcanics mark the top of the sequence and are represented by amygdaloidal pillow and pillow breccia flows. Pillow tops indicate younging to the east and Slimmon (this volume) states that the rocks west of Table Lake also young to the east. The observed stratigraphy suggests that this ultramafic-mafic rock sequence may represent part of a remnant ophiolitic succession.

The entire sequence is variably strained. Schistose and/or talcose rocks are found within the ultramafic intrusions. The mafic intrusions, sill/dyke complex and mafic volcanic rocks exhibit localized zones of millimetre-scale tectono-metamorphic mylonitic layering. Contacts between the various units are typically marked by topographic low lineaments commonly with steep-sided ridges of sheared rocks which are interpreted as shear zones which dismember the sequence. Kinematic indicators are scarce but appear to indicate west-side-up thrusting which is consistent with the vergence of imbrication in the region.

d) Mafic-felsic Pyroclastic Volcanic Rock Sequence

A mafic-felsic pyroclastic volcanic rock sequence crops out in the area between Barney Lake and Tanyss Lake (Figure 1), overlying the ultramafic-mafic rock sequence to the west. The sequence strikes north-south and dips steeply to the west. From the west to east, the inferred younging direction, the base of the pyroclastic sequence consists of a mafic crystal/ash tuff. This unit is overlain by a mafic monolithic tuff breccia and topped by a mafic heterolithic tuff breccia. Fragments in the heterolithic tuff breccia are predominantly mafic to intermediate in composition, however, felsic fragments are found locally. Felsic volcanic rocks are subordinate in the sequence and are typically about 100 m thick. These are dominated by a monolithic tuff breccia with interlayered quartz crystal tuff. The felsic volcanic rocks are found interlayered within the mafic tuff breccias. Silicification is locally found associated with the felsic volcanic rocks. The differentiation of primary felsic rocks and silicified mafic rocks can be difficult. Contacts within the pyroclastic sequence appear to be transitional and units are commonly interlayered.

e) Reynard Lake Pluton

The Reynard Lake Pluton is a complex multiphase intrusion which extends over 30 km from the Amy-Batty Lakes area north of the map area to Tanyss Lake in the south (Figure 1) and is approximately 5 km wide. A single-zircon Pb age date of 1853 ±8 Ma was obtained recently from a granite phase located in the central portion of the intrusion (Ansdell and Kyser, 1991). At least nine components were recognized and mapped during this field season. Contacts between these phases range from gradational to sharp and cross-cutting. Although the relative timing of emplacement of these components is not everywhere clear, there appears to be an evolution from early mafic phases to later more felsic phases. A summary of distinct phases is listed below.

1. Mesocratic to melanocratic gabbro locally exhibits primary igneous layering defined by hornblende-rich and plagioclase-rich bands. Layering is accentuated by a strong foliation which appears gneissic in places. Mafic bands are in places strongly magnetic.

2. Mesocratic to leucocratic gabbro/diorite also exhibits local igneous layering and a strong foliation.

3. Plagioclase porphyritic, weakly to moderately foliated, medium- to coarse-grained diorite. Pegmatitic phases are observed locally.

4. Fine- to medium-grained, weakly to moderately foliated diorite that is similar to and gradational with the coarser-grained plagioclase porphyritic diorite.

5. Fine- to medium-grained, weakly to moderately foliated quartz diorite/diorite.

6. Medium-grained weakly to moderately foliated hornblende granodiorite.

7. Medium- to coarse-grained, sheared quartz-eye granite/granodiorite.

8. Mixed rock dominated by leucogabbro/diorite, hornblende porphyritic tonalite and quartz porphyritic granite/granodiorite. Other intrusive components are present locally and intrusion breccias are common.

9. Fine- to medium-grained, typically plagioclase porphyritic, massive gabbro is a marginal phase generally occurring as sills and dykes.

A prominent aureole of contact strain surrounds the Reynard Lake Pluton in a zone up to 100 m in width in which the early D1 foliation is folded. The mesoscopic folds (F2) in the aureole are open to closed and an axial planar cleavage (S2) is developed locally. Fold axes plunge steeply in the aureole and are coaxial with a well-developed stretching lineation (L2), typically expressed as an L-tectonite, in both the supracrustal rocks and the intrusion. The intersection of the F2 axial planes with the S1 foliation has produced an intersection lineation which is also coaxial with L2. The strong linear mesoscopic structures found within the contact strain aureole of the Reynard Lake Pluton are attributed to its diapiric emplacement.

4. Mineral Potential

a) Base Metals

Bailes and Syme (1987) noted that most massive sulphide deposits in the Flin Flon-White Lake area are stratigraphically associated with large felsic volcanic complexes in direct or indirect association with sediments and volcaniclastic rocks. The felsic volcanic units identified in the mafic-felsic pyroclastic sequence during the past summer, although of limited width and strike length, warrant further evaluation.

The Coronation Mine, located on the east side of Phil Lake (Figure 1), represents the only economic deposit found to date in the immediate map area. Between 1960 and 1964 approximately 1,412,860 tons of ore were mined yielding 4.26 percent Cu, 0.15 percent Zn, 4.68 g/ton Ag and 1.87 g/ton Au (Franklin and Thorpe, 1982). A collection of papers on various aspects of the geology, geophysics and geochemistry of the Coronation Mine was published by the Geological Survey of Canada (Byers, 1968). A description of the mine is also found in Byers *et al.* (1965). Unfortunately, the only outcrops in the vicinity of the mine are found east of the zone of mineralization and consist of massive, amygdaloidal pillowed and pillow breccia mafic volcanics. The Coronation Mine ore bodies are hosted by these mafic volcanic rocks which strike northwest, dip steeply to the southwest and have a stretching lineation which plunges moderately to steeply to the south. The deposit consists of three ore shoots within a zone of lower grade mineralization which is 275 m long, up to 37 m wide and extends to a vertical depth of 305 m. Chalcopyrite is the only significant ore mineral. The three ore shoots and mineralized zone plunge parallel to the stretching lineation in the volcanics. Froese (1968) considered cordierite-anthophyllite found associated with the ore bodies to be regionally metamorphosed zones of chloritization presumably produced at the time of mineralization prior to metamorphism.

The mafic volcanic rocks which host the ore bodies are characterized by oceanic, rift-type geochemical features (Parslow and Gaskarth, 1988) and are located at the top of the ultramafic-mafic rock sequence which may represent a remnant ophiolitic sequence. These characteristics are consistent with a Cyprus-type deposit.

b) Gold

Two phases of gold mineralization have been recognized by Galley and Franklin (1987) in the Phantom Lake area to the north of the map area: 1) an early phase of porphyry-type mineralization and alteration associated with the Boot Lake-Phantom Lake intrusions; and 2) a later phase of mineralization and alteration controlled by brittle-ductile shear zones related to the Cliff Lake-Ross Lake Fault System which is believed to have developed during a protracted event. Concentrations of gold with the greatest economic potential have been found so far in the brittle-ductile shear zones.

Zones of high strain were produced during four stages of the deformational history of the map area (see Reilly, 1990) as follows:

D1: Ductile structures dominate in broad shear zones (i.e. Mystic Lake Shear Zone);

D2: Predominantly ductile structures in contact strain aureoles of the Reynard Lake Pluton and the Kaminis Lake Intrusion;

D3: Discrete brittle-ductile shear zones which demarcate the transition from a ductile regime to a brittle regime; and

D4: Brittle structures associated with faulting (i.e Mystic Lake Fault).

The high strain zones in the map area are commonly chloritized, silicified, sericitized, carbonatized, hematized and are locally gossanous with sulphide mineralization. D3 brittle-ductile shear zones are perhaps the most favorable targets for gold mineralization.

c) Platinum Group Elements

Prior to the Namew Lake discovery the only nickel sulphides in the Flin Flon area were found as weak disseminations in serpentinized ultramafic rocks or along the contacts between relatively unaltered mafic-ultramafic intrusions and volcanogenic sulphide-rich country rock (Pickell, 1987). The Namew Lake nickel-copper deposit lies 60 km south of Flin Flon. The orebody mineralization is concentrated around the top of a pipe-like sill of pyroxenite-peridotite near the interface between magnetite-bearing diorite sills and aplitic quartz monzonite gneisses. The ultramafic sill is late and mantle-derived, with komatiitic affinities and schistose and/or talcose margins (Pickell, 1987). Ultramafic sills found in the map area are schistose and/or talcose locally. Although no sulphides where observed, these sills may warrant further examination for Ni-Cu-PGE mineralization in light of the Namew-type magmatic sulphide deposit.

5. References

Alcock, F.J. (1923): Flin Flon map area, Manitoba and Saskatchewan; Geol. Surv. Can., Summ. Rep. 1922, ptC, p1-36.

Ambrose, J.W. (1936a): Progressive kinetic metamorphism in the Missi Series near Flin Flon, Manitoba; Am. J. Sci. (5th ser.), v32, p257-286.

__________ (1936b): Structures in the Missi Series near Flin Flon, Manitoba; Trans. R. Soc. Can., (3rd ser.), v30, sec4, p81-98.

Ansdell, K.M. and Kyser, K.T. (1991): Plutonism, deformation and metamorphism in the Proterozoic Flin Flon greenstone belt, Canada: Limits on timing provided by the single-zircon Pb-evaporation technique; Geol., v19, p.518-521.

Bailes, A.H. and Syme, E.C. (1987): Geology of the Flin Flon-White Lake area; Man. Energy Mines, Geol. Rep. GR 87-1, 313p.

Bruce, E.L. (1916): Amisk-Athapapuskow Lake area, northern Saskatchewan and northern Manitoba; Geol. Surv. Can., Sum. Rep. 1915, p126-130.

__________ (1918a): Schist Lake district, northern Manitoba; Geol. Surv. Can., Sum. Rep. 1917, ptD.

__________ (1918b): Amisk - Athapapuskow Lake district; Geol. Surv. Can., Mem. 105, 91p.

Byers, A.R. ed. (1968): Symposium on the Geology of the Coronation Mine; Geol. Surv. Can., Pap. 68-5, 329p.

Byers, A.R. and Dahlstrom, C.D.A. (1954): Geology and mineral deposits of the Amisk-Wildnest Lakes area, 63L-9, 63L-16, Saskatchewan; Sask. Dep. Miner. Resour., Rep. 14, 177p.

Byers, A.R., Kirkland S.J.T. and Pearson, W.J. (1965): Geology and mineral deposits of the Flin Flon area, Saskatchewan; Sask. Dep. Miner. Resour., Rep. 62, 95p.

Franklin, J.M. and Thorpe, R.I. (1982): Comparative metallogeny of the Superior, Slave and Churchill Provinces; *in* Hutchinson, R.W., Spence, C.D. and Franklin, J.M. (eds.), Precambrian Sulphide Deposits, Geol. Assoc. Can., Spec. Pap. 25, p3-90.

Froese, E. (1968): Metamorphic rocks from the Coronation Mine and surrounding area; *in* Byers, A.R. (ed.), Symposium on the geology of the Coronation Mine, Geol. Surv. Can., Pap. 68-5, p55-77.

Galley, A.G. and Franklin, J.M. (1987): Geological setting of gold , copper, tungsten and molybdenum occurrences in the Phantom Lake region; *in* Summary of Investigations 1987, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 87-4, p115-124.

Gordon, T.M., Hunt, P.A., Loveridge, W.D., Bailes, A.H. and Syme, E.C. (1987): U-Pb zircon ages from the Flin Flon and Kisseynew belts, Manitoba: chronology of Early Proterozoic crust formation; Geol. Assoc. Can./Miner. Assoc. Can., Joint Annual Meeting, Prog. Abstr., v12, p48.

Hoffman, P.F. (1990): Subdivision of the Churchill Province and extent of the Trans-Hudson Orogen; *in* Lewry, J.F. and Stauffer, M.R. (eds.), The Early Proterozoic Trans-Hudson Orogen of North America, Geol. Assoc. Can., Spec. Pap. 37, p15-39.

Heywood, W.W. (1966): Ledge Lake area, Manitoba and Saskatchewan; Geol. Surv. Can., Mem. 337, 43p.

Lewry, J.F. and Collerson, K.D. (1990): The Trans-Hudson Orogen: Extent, subdivision, and problems; *in* Lewry, J.F. and Stauffer, M.R. (eds.), The Early Proterozoic Trans-Hudson Orogen of North America: Geol. Assoc. Can., Spec. Pap. 37, p1-14.

MacQuarrie, R. (1980): Absolute age of the Flin Flon deposit; abstr., CIM Bull. v73, no820, p53.

McInnes, W. (1909): Explorations on the Churchill River and South Indian Lake; Geol. Surv. Can., Summ. Rep., 1908, p87-92.

Parslow, G.R. and Gaskarth, J.W. (1988): Proterozoic rocks of east-central Saskatchewan: Geochemistry, structure and mineralization controls; *in* Summary of Investigations 1988, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 88-4, p127-139.

Pickell, J.R. (1987): The Namew Lake nickel-copper discovery, Flin Flon, Manitoba; Geol. Assoc. Can./Miner. Assoc. Can., Joint Annual Meeting, Prog. Abstr. v12, p80.

Reilly, B.A. (1990): Bedrock geological mapping, Mystic Lake-West Arm, Schist Lake area (part of NTS 63K-12); *in* Summary of Investigations 1990, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 90-4, p25-35.

Stockwell, C.H. (1946): Flin Flon-Mandy area, Manitoba and Saskatchewan; Geol. Surv. Can., Pap. 46-14, 5p.

__________ (1960): Flin Flon-Mandy area, Manitoba and Saskatchewan; Geol Surv. Can., Map 1078A (with marginal notes), scale 1:12,000.

Syme, E.C. (1985): Athapapuskow Lake Project; *in* Report of Field Activities 1985, Man. Energy Mines, Misc. Rep., p44-47.

__________ (1986): Schist Lake area, Athapapuskow Project; *in* Report of Field Activities 1986, Man. Energy Mines, Misc. Rep., p36-42.

__________ (1987): Athapapuskow Lake Project; *in* Report of Field Activities 1987, Man. Energy Mines, Misc. Rep., p30-39.

__________ (1988): Athapapuskow Lake Project; *in* Report of Field Activities 1988, Man. Energy Mines, Misc. Rep., p20-34.

Syme, E.C., Bailes, A.H., Gordon, T.M. and Hunt, P.A. (1987): U-Pb zircon geochronology in the Flin Flon belt: age of Amisk volcanism; *in* Report of Field Activities 1987, Man. Energy Mines, p105-107.

Tanton, T.L. (1941a): Flin Flon, Saskatchewan and Manitoba; Geol. Surv. Can., Map 632A (with marginal notes), scale 1:63,360.

__________ (1941b): Schist Lake, Saskatchewan-Manitoba; Geol. Surv. Can., Map 633A (with marginal notes), scale 1:63,360.

Thomas, D.J. (1989): Geology of the Douglas Lake-Phantom Lake area (part of NTS 63K-12 and -13); *in* Summary of Investigations 1989, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 89-4, p44-54.

__________ (1990): New perspectives of the Amisk Group and regional metallogeny, Douglas Lake-Phantom Lake Area; *in* Summary of Investigations 1990, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 90-4, p13-20.

Tyrrell, J.B. (1902): Explorations in the northeastern portion of the district of Saskatchewan and adjacent parts of the District of Keewatin; Geol. Surv. Can., Ann. Rep., v8, Rep. F, 48p.

Revision Bedrock Geological Mapping, Table Lake Area (Part of NTS 63L-9)[1]

W.L. Slimmon

Slimmon, W.L. (1991): Revision bedrock geological mapping, Table Lake area (part of NTS 63L-9); *in* Summary of Investigations 1991, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 91-4.

This project is part of the Manitoba-Saskatchewan Shield Margin NATMAP Project of multi-disciplinary geoscience studies along the Precambrian Shield-Paleozoic margin. In Saskatchewan, bedrock geological mapping is part of 'Project Seagull', which also includes mapping by Thomas (1989, 1990, this volume) and Reilly (1990, this volume). The Table Lake area, which was remapped this summer at a scale of 1:12 500, is situated 20 km southwest of Flin Flon and covers approximately 96 km^2 north of the Shield margin between Amisk Lake and the east shore of Table Lake (Figure 1).

Surface topography is quite flat throughout the area and has a maximum relief of less than 25 m. Along the shore of Amisk Lake extensive stretches contain up to 90 percent bedrock exposure. Inland the exposure is about 25 percent in the north half of the map area, and less than 5 percent in the south half where isolated outcrops are interspersed between expansive open and wooded swamps.

Previous bedrock geological surveys in the area are reported by McInnes (1913), Bruce (1918), Wright (1933), Mawdsley (1934), Byers and Dahlstrom (1954), and McDougall *et al.* (1979). Quaternary geology was interpreted by Schreiner (1984).

Geophysical aeromagnetic surveys of the area comprise Geological Survey of Canada 1:250 000 scale map 7754G (63L) and 1:63 000 scale map 4606G (63L-9), and more recent high resolution vertical gradient and total field 1:20 000 scale maps included in GSC Open File 1626.

Geochemical surveys encompassing the area include regional lake sediment geochemical surveys undertaken by Hornbrook *et al.* (1977 and 1985), and a regional lithogeochemical survey by Parslow and Gaskarth (1984).

1. Regional Geology

The Table Lake area is situated in the Flin Flon-Snow Lake Domain which comprises a relatively low metamorphic grade component of the Early Proterozoic Trans-Hudson Orogen. There is a transitional boundary to the north with high-grade gneisses of the Kisseynew Domain, and the domain is overlain to the south by flat-lying unmetamorphosed Ordovician dolomites and dolomitic limestones of the Red River Formation, and marine sandstones of the Winnipeg Formation.

In the Flin Flon-Snow Lake Domain, the Amisk Group has been described as predominantly comprising subaqueous mafic flow related rocks, subordinate felsic flow related rocks and porphyries, and minor amounts of subaerial volcaniclastic rocks, sedimentary rocks, and syn- to post-volcanic dykes and sills. The group is unconformably overlain by molasse sediments of the Missi Group and intruded by a variety of major pre- to post-Amisk Group and post-Missi Group intrusions. U-Pb zircon geochronology indicates Amisk Group volcanism at ca. 1886 Ma (Syme *et al.*, 1987) and Missi Group deposition at ca. 1832 Ma (Gordon *et al.*, 1987).

In the Flin Flon-East Amisk Lake region, the Amisk Group has more recently been recognized as comprising a collage of volcanic assemblages from different tectonomagmatic environments defined by the geochemical and physical nature of the rocks. In the Flin Flon-Phantom Lake area the rocks are tholeiitic and described as having an oceanic island arc affinity (Gaskarth and Parslow, 1987; Parslow and Gaskarth, 1988; Thomas, 1990; Watters, 1989 and 1990). In contrast, the East Amisk Lake rocks have an overlapping tholeiitic MORB and within-plate chemical signature which is interpreted to indicate either a MORB, ophiolitic, ocean island or back-arc basin rift origin (Parslow and Gaskarth, 1984 and 1988; Gaskarth and Parslow, 1987). Mapping by Thomas (1989 and 1990) and Reilly (1990) indicates that the two contrasting tectonomagmatic environments are separated by a tectonic contact zone of alteration and brittle-ductile shearing, tentatively interpreted to result from northeastward thrusting (Thomas *et al.*, 1991).

2. General Geology

Approximately 75 percent of the exposed portion of the Table Lake area is underlain by mafic metavolcanic rocks. Due to the poor exposure caused by lichen and moss cover on the outcrops, primary features in the metavolcanics are best seen along the shore of Amisk Lake. Because of this poor exposure, much of the mafic metavolcanic rocks, located farther inland, have not been subdivided. Where primary features are observed,

---

(1) Saskatchewan project A.113, from which this report derives, was funded in 1991 under the Canada-Saskatchewan Partnership Agreement on Mineral Development 1990-95.

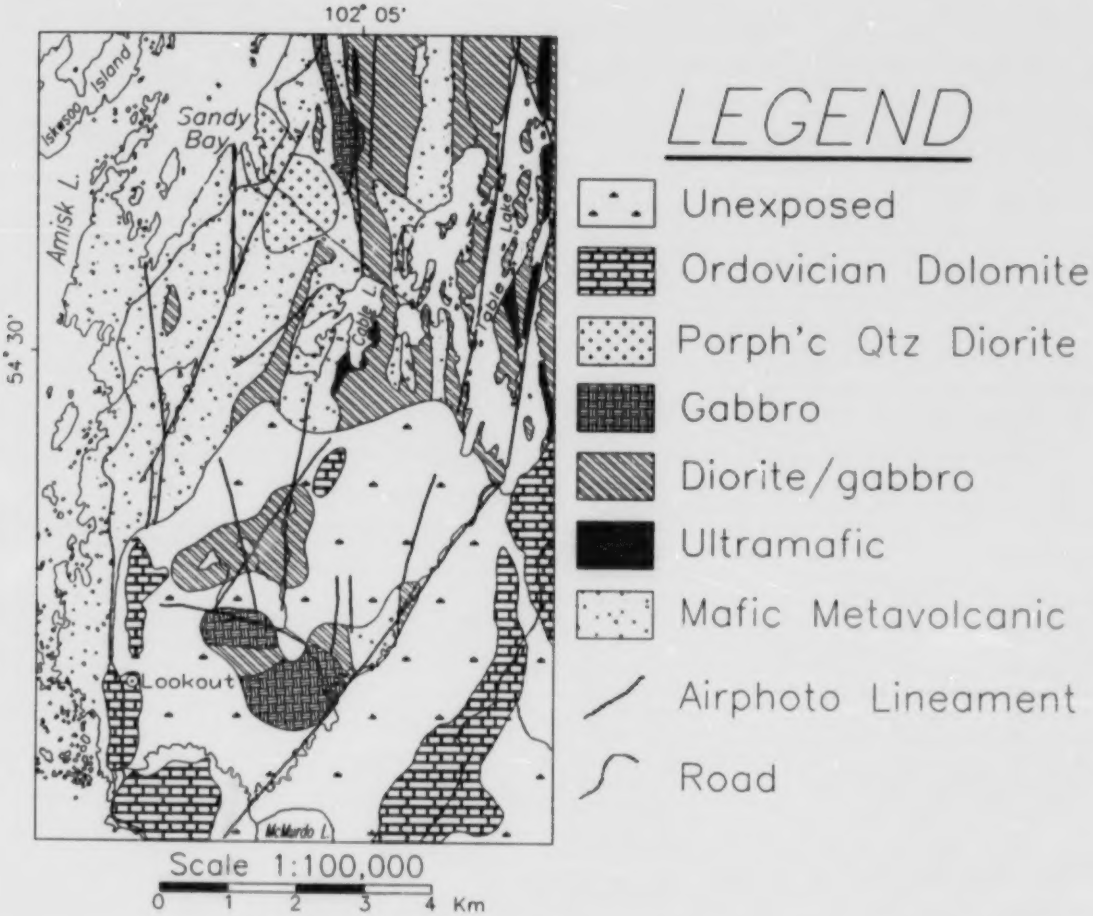


Figure 1-General geology of the Table Lake area.

the mafic metavolcanic rocks comprise mostly pillowed and massive flows with minor pillow and flow breccias.

Major intrusive rocks, forming north-south-trending linear belts up to 2 km wide, are primarily dioritic to gabbroic, but vary from quartz diorite to ultramafic. These large sill-like bodies intrude the metavolcanic rocks and have subconcordant contacts with them. The gabbroic rocks have virtually the same chemical composition as the metavolcanic rocks and are considered to be their intrusive equivalent (Parslow and Gaskarth, 1984). Much of the unexposed area southwest of Table Lake is interpreted on the basis of geophysics to be underlain by gabbro. Younger stock-like quartz-feldspar porphyritic quartz diorite intrusions up to 1.5 km in diameter cut both the metavolcanic rocks and the gabbros.

Minor intrusive rocks which intrude the metavolcanic rocks include gabbroic dykes and sills, feldspar porphyry dykes and rare felsite dykes. Rare diabase dykes intrude the gabbroic rocks around Table Lake. Quartz-feldspar porphyry dykes, the youngest dykes in the area, are associated with the larger quartz-feldspar porphyritic quartz diorite stocks and radiate from them to intrude both the metavolcanic and gabbroic rocks.

3. Lithological Units

a) Mafic Metavolcanic Rocks

Between Amisk Lake and Cable Lake, mafic metavolcanic rocks comprise a continuous sequence of flow related basaltic rocks nearly 4.5 km thick. The rocks are typically fine grained with bluish grey fresh surfaces that weather greenish grey. Quartz and calcite stringers are common and locally the rocks contain vuggy quartz veins up to 15 cm wide. Primary minerals have been extensively altered to a felted aggregate containing variable amounts of hornblende, actinolite, chlorite, biotite, plagioclase, epidote, sericite and carbonate. Accessory minerals include quartz, apatite, magnetite, pyrite and pyrrhotite. A schistose foliation is absent or only very weakly developed in most of these rocks, but in areas close to structural lineaments and along the shore region of Amisk Lake a stronger fabric occurs. Weak to moderate silicification is common along the Amisk Lake shoreline, and in the vicinity of Sandy Bay the rocks are strongly silicified and have the appearance of felsic metavolcanic rocks. Epidotization is relatively common and occurs throughout the mafic metavolcanic rocks as cross-cutting epidote stringers and in pods up to a few centimetres wide.

Pillow flows (unit Smp) are common in a 1 to 2 km wide zone along the shore of Amisk Lake. The pillows range from 20 to 200 cm in length, with an average of 50 cm, and have a typical length to width ratio of about 2 to 1 in the Y-Z plane of the finite strain ellipsoid. This ratio is as much as 20 to 1 in local high strain zones. The pillows are most commonly aphyric and non-amygdaloidal, although in places **amygdaloidal pillow flows (unit Smpa)** exist which contain sparse 2 to 3 mm size calcite amygdules, typically concentrated around the pillow margins, and in some locations as radial amygdules. Rare **feldspar phyric pillow flows (unit Smpf)** contain less than 5 percent anhedral to euhedral feldspar phenocrysts up to 5 mm long. Pillow selvages are typically less than 5 mm wide and are more melanocratic than the pillows. The pillows are generally more amygdaloidal and have wider selvages in the north. Several of the pillow flow exposures display a consistent eastward facing direction.

Pillow/flow breccias (unit Smb) occur locally as narrow horizons intercalated with the pillow flows, and as more extensive units northwest and southwest of Cable Lake. Mafic, unsorted monogenetic clasts up to several centimetres in size are clast supported. The matrix is typically a more melanocratic basaltic rock, but locally it consists of carbonates.

Feldspar phyric flows (unit Smf) occur only in a small area southeast of Sandy Bay. The rocks contain up to 5 percent anhedral to euhedral feldspar phenocrysts less than 5 mm in length.

The mafic metavolcanic rocks sandwiched between the gabbroic intrusions northwest of Table Lake are distinctly different than those just described, being much darker, more amphibolitic, and extensively networked by cross-cutting feldspar stringers and veins. Silicification is absent and epidotization is rare.

b) Major Intrusive Rocks

Quartz-feldspar porphyritic quartz diorite (unit qd) outcrops east of Sandy Bay and in two smaller bodies west of Cable Lake and Table Lake. This unit intrudes the metavolcanic and gabbroic rocks. Xenoliths of the country rock are common in marginal zones. The rock is generally medium grained and varies from light grey to greyish-orange pink. Groundmass minerals consist of extensively saussuritized plagioclase, 10 percent quartz, 5 percent chlorite psuedomorphs of biotite, 2 to 3 percent epidote, and traces of accessory apatite, sphene, zircon, pyrite and magnetite. The size and amount of quartz and plagioclase phenocrysts varies locally. The Sandy Bay exposure contains 5 to 10 percent round to oval quartz phenocrysts less than 5 mm in diameter; large plagioclase phenocrysts are not prominent. The Cable Lake and Table Lake exposures contain 5 to 15 percent amoeboid-shaped quartz aggregates up to 10 mm in diameter with as much as 35 percent euhedral plagioclase phenocrysts up tc 20 mm long.

A tectonic fabric is either absent or locally very weakly displayed by a slight flattening of the quartz phenocrysts.

Quartz diorite (unit Tqd) is associated with the gabbroic rocks at Table Lake and north of Meridian Creek. The typically weakly-foliated to massive, light-grey, fine-grained rock contains an average of 10 percent quartz and 30 percent hornblende. Cross-cutting feldspar and epidote veins occur locally.

Diorite/gabbro (unit Tdg) occurs as north-south-trending, sill-like bodies in the Table Lake area. These leucocratic to mesocratic, fine-grained and generally massive- to weakly-foliated rocks are more strongly foliated along structural lineaments that pass through Table Lake. Locally the outcrops contain centimetre scale feldspar sweats and cross-cutting veins. In thin section the rocks are extensively altered and comprise saussuritized plagioclase and up to 40 percent actinolite pseudomorphs after hornblende.

Gabbro (unit Tg) underlies an area northwest of Table Lake and another area north of Meridian Creek. The rocks are medium grained with a massive, equigranular texture, comprising plagioclase, 25 to 40 percent hornblende psuedomorphs after pyroxene, some with pyroxene cores, and up to 5 percent magnetite. Northwest of Table Lake the plagioclase is strongly saussuritized. In the Meridian Creek exposure the plagioclase is relatively unaltered, suggesting that it is a somewhat later intrusion. The Meridian Creek outcrops are highly magnetic and contain mafic metavolcanic xenoliths as well as up to 5 percent medium- to coarse-grained, 10 cm wide pegmatite veins.

Ultramafic rocks of primarily pyroxenitic to peridotitic composition exist in 100 to 200 m wide, sill-like exposures along the west side of Table Lake and at the southeast corner of Cable Lake. A highly serpentinized, **fine-grained ultramafic (unit Ts)** underlies a relatively undeformed, less-altered, **medium-grained ultramafic (unit Tp)** at both locations. The highly serpentinized variety is possibly sheared and contains talc as well as remnant pyroxenes. Both rock types contain up to 10 percent magnetite. Although more common in the serpentinized variety, both types are locally cross-cut by chrysotile stringers and veinlets typically less than 1 cm wide, but in places up to 5 cm wide.

These ultramafic rocks possibly comprise the lower part of a dismembered ophiolite sequence that is further exposed east of Table Lake (see Reilly, this volume).

c) Minor Intrusive Rocks

Quartz-feldspar porphyry dykes have the same composition as the porphyritic quartz diorite stocks (unit qd) and radiate from them in all directions. They are concentrated within a 200 to 300 m envelope around the stocks; however, they also exist farther away. The dykes range from a few centimetres to several metres in width and average about 2 m wide.

Feldspar porphyry dykes intrude the metavolcanic rocks between Amisk Lake and Cable Lake but are absent in those northwest of Table Lake. These dykes average 3 m wide and have a maximum width of 12 m. They are most commonly subconcordant to the metavolcanic rocks in a north-trending orientation, but several are strongly discordant and have northeast to east orientations. The dykes are dacitic and contain 10 to 30 percent anhedral to euhedral plagioclase phenocrysts averaging 5 mm long.

Felsite dykes are relatively rare and intrude the metavolcanic rocks with a dominant north-trending subconcordant orientation. These light grey aphanitic rocks are dacitic in composition and average 1 m in width.

Gabbroic dykes and sills up to 100 m wide have subconcordant locally sheared boundaries with the metavolcanic rocks. These fine- to medium-grained rocks contain 30 percent hornblende and commonly 1 to 2 percent pyrite.

Diabase dykes, averaging 2 m wide, are present in only three locations and intrude both the metavolcanic and gabbroic rocks at the south end of Table Lake.

4. Structural Geology

The earliest recognizable structure in the map area is a weakly developed, north-south-striking, schistose foliation. The foliation has a moderate to steep, predominantly west dip, although in the Sandy Bay area the dip is very steep and varies from east to west. A moderately-plunging, west-trending stretching lineation accompanies the foliation in a few localities along the east shore of Table Lake. Pillow structures and various other volcanic features indicate that the rocks have been overturned and young to the east.

Local brittle-ductile shear zones from less than a metre to several metres wide are subconcordant to the north-striking foliation and are well exposed along the shore of Amisk Lake. In the shear zones, pillows are stretched in the Y-Z plane of the finite strain ellipsoid to ratios of 20 to 1. In the southern part of the map area along Amisk Lake, the shear zones are locally accompanied by a moderately south-southeast-plunging stretching lineation that deforms pillows and flow breccia clasts. These shear zones are possibly related to north-trending regional scale lineaments, where similar shearing occurs locally. Kinematically, the small-scale shears and one of the large-scale structures that cuts the intrusive stock east of Sandy Bay have a dominant dextral strike slip as determined from drag folds and lithology boundary offsets.

Brittle shearing is rare, although one very fissile and extensively chloritized zone up to 10 m wide occurs near the top of a ridge along the east shore of Sandy Bay. This north-trending zone cuts mafic metavolcanic rocks less than 200 m from the metavolcanic-Sandy Bay quartz diorite contact.

5. Metamorphism

Mineral assemblages and psuedomorphs indicate that the Table Lake area attained lower amphibolite metamorphic facies, but many of the rocks have retrogressed to greenschist facies. In general, the dark amphibolitic metavolcanic rocks northwest of Table Lake and the medium-grained gabbros and ultramafic rocks have remained at amphibolite grade, whereas the lighter greenish metavolcanics rocks east of Cable Lake and the fine-grained diorite/gabbros and serpentinized ultramafic rocks have retrogressed to greenschist grade. The quartz-feldspar porphyritic quartz diorite intrusions are at greenschist facies and there is no indication that they attained amphibolite facies.

6. Economic Geology

Exploration dates back to the early 1900's and 1930's when prospecting for gold was carried out in the Amisk Lake area. In the early 1950s exploration concentrated on the search for base metals and several diamond drill holes were completed to test a number of electromagnetic conductors. The north-south-trending conductors are primarily located beneath Table Lake, in the Cable Lake area and along the east shore of Amisk Lake and occur in both the metavolcanic and intrusive rocks. The drill holes intersected several massive sulphide zones comprising pyrite, pyrrhotite and minor chalcopyrite which variably assay up to 0.15 percent Cu, 0.19 percent Zn, 0.24 percent Ni, 0.12 percent Ag and 0.01 percent Au over intervals of less than a metre (Saskatchewan Energy and Mines Mineral Deposits Index).

In the late 1980s the area was again prospected for gold and grab samples from the southwest part of the map area assayed up to 0.014 oz/ton Au (Saskatchewan Energy and Mines Assessment File 63L9-0343).

No new extensively mineralized zones were discovered during mapping this summer. The shear zone along the east shore of Sandy Bay locally contains 2 to 3 percent pyrite in the metavolcanic wall rocks. Ground EM surveys performed in 1954 and 1974 indicate no conductivity over the shear zone. There is no indication of previous work on the shear zone itself and the area is currently open.

7. References

Bruce, E.L. (1918): Amisk-Athapapuskow Lake District; Geol. Surv. Can., Mem. 105, 91p.

Byers, A.R. and Dahlstrom, C.D.A. (1954): Geology and mineral deposits of the Amisk-Wildnest Lakes area, Saskatchewan; Sask. Dep. Miner. Resour., Rep. 14, 177p.

Gaskarth, J.W. and Parslow, G.R. (1987): Proterozoic volcanism in the Flin Flon greenstone belt, east-central Saskatchewan, Canada; *in* Geochemistry and Mineralization of Proterozoic Volcanic Suites; Geol. Soc. Lon., Spec. Publ. 33, p183-200.

Gordon, T.M., Hunt, P.A., Loveridge, W.D., Bailes, A.H. and Syme, E.C. (1987): U-Pb zircon ages from the Flin Flon and Kisseynew belts, Manitoba: chronology of Early Proterozoic crust formation; GAC-MAC Joint Annual Meeting, May 1987, Saskatoon, Prog. Abstr., v12, p48.

Hornbrook, E.H.W., Garrett, R.G., Lynch, J.J. and Beck, L.S. (1977): Regional lake sediment geochemical reconnaissance data, east-central Saskatchewan (NTS 63M, 64D and parts of 63K, L, N, 73I, O, P and 74A); Geol. Surv. Can., Open File 488.

Hornbrook, E.H.W., Lynch, J.J., Friske, P.W.B., Lund, N.D. and Schmitt, H.R. (1985): Regional lake sediment geochemical reconnaissance data, east-central Saskatchewan (NTS 63M, 64D and parts of 63K, L,N, 73I, O, P and 74A); Geol. Surv. Can., Open File 1129, 150p.

Mawdsley, J.B. (1934): A brief outline of the geological history of the northwest shore of Lac la Ronge and the Beaver Lake areas; Sask. Dep. Nat. Resour., Annu. Rep. (1933), App. 2, 20p.

McDougall, F.H., Parslow, G.R. and Watters, B.R. (1979): Flin Flon base metals project; *in* Summary of Investigations 1979, Saskatchewan Geological Survey, Sask. Miner. Resour., Misc. Rep. 79-10, p116-119.

McInnes, W. (1913): The basins of the Nelson and Churchill Rivers; Geol. Surv. Can., Mem. 30, 146p.

Parslow, G.R. and Gaskarth, J.W. (1984): Geochemistry of the east Amisk area; Sask. Energy Mines, Open File Rep. 84-23, 156p.

__________ (1988): Proterozoic rocks of east-central Saskatchewan: geochemistry, structure and mineralization controls; *in* Summary of Investigations 1988, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 88-4, p127-139.

Reilly, B.A. (1990): Bedrock geological mapping, Mystic-West Arm, Schist Lakes area (northwest part of NTS 63K-12); *in* Summary of Investigations 1990, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 90-4, p25-35.

Schreiner, B.T. (1984): Quaternary geology of the Amisk Lake area (NTS 63K, L); Sask. Energy Mines, Open File Rep. 84-2.

Syme, E.C., Bailes, A.H., Godon, T.M. and Hunt, P.A. (1987): U-Pb zircon geochronology in the Flin Flon belt: age of Amisk volcanism; *in* Report of Field Activities, 1987, Manit. Energy Mines, Misc. Rep., p105-107.

Thomas, D.J. (1989): Geology of the Douglas Lake-Phantom Lake area (part of NTS 63K-12 and -13); *in* Summary of Investigations 1989, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 89-4, p44-54.

__________ (1990): New perspective on the Amisk Group and regional metallogeny, Douglas Lake-Phantom Lake area, northern Saskatchewan; *in* Summary of Investigations 1990, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 90-4, p13-20.

Thomas, D.J., Reilly, B.A., Slimmon, W.L. and Watters, B.R. (1991): Tectonostratigraphic controls on massive sulphide deposits in the Amisk Group: Flin Flon-Amisk Lake area; The Geological Society of CIM, First Annual Field Conference, Saskatoon, Sept. 1991, Prog. Abstr., p26.

Watters, B.R. (1989): Lithogeochemical studies in the Flin Flon-Amisk Lake area and Glennie Domain; *in* Summary of Investigations 1989, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 89-4, p55-57.

__________ (1990): Lithogeochemical studies in the Flin Flon-Amisk Lake area; *in* Summary of Investigations 1990, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 90-4, p44-47.

Wright, J.F. (1933): Amisk Lake area, Saskatchewan; Geol. Surv. Can., Summ. Rep. 1932, ptC, p73-110.

Lithogeochemical Studies in the East Amisk Lake Area[1]

B.R. Watters[2]

Watters, B.R. (1991): Lithogeochemical studies in the East Amisk Lake area; *in* Summary of Investigations 1991, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 91-4.

A lithogeochemical study of metavolcanic and intrusive rock units in the Flin Flon-Amisk Lake area continued during 1991. This study is being carried out in collaboration with the detailed bedrock remapping program being conducted in the Phantom Lake-Douglas Lake area (Thomas, 1989, 1990, this volume), the Mystic Lake-West Arm, Schist Lake area (Reilly, 1990, this volume) and the East Amisk area (Slimmon, this volume). The following brief report deals with some aspects of the study only and includes results obtained prior to the 1991 field season.

1. Progress to Date

Major and trace element data have been obtained for approximately 300 samples from the Douglas Lake-Phantom Lake-Birch Lake area (Thomas, 1989, 1990, this volume) and from the Mystic Lake-West Arm, Schist Lake area (Reilly, 1990, this volume). Sample suites analysed are mainly volcanic rock types but representatives of the main intrusions were also sampled. A further 170 samples from these areas as well as from the East Amisk area (Slimmon, this volume), collected during the 1991 field season, are presently being analysed.

A limited suite of 38 samples of metavolcanic rocks from the Snake Rapids area (Ashton, 1990) have also been analysed and a brief discussion of results from that study is presented separately (Watters and Ashton, this volume).

2. Metavolcanic Rocks

Descriptions of the metavolcanic sequences in the areas under discussion have been provided by Thomas (1989, 1990) and Reilly (1990).

Metavolcanic rocks from the Douglas Lake-Phantom Lake area are sub-alkaline in nature and range in composition from basaltic to rhyolitic with a dominance of basic rocks (Figure 1). Intermediate types are less common than rhyolite, resulting in a near bimodal distribution of rock types, an observation consistent with the field data (Thomas, 1989). As reported in Watters (1989), it can be demonstrated that silica has been mobile to varying degrees in these rock units and, therefore, the commonly used classification schemes based on that element can produce ambiguous results. In Figure 1 the elements Zr, Ti, Nb and Y, all of which are considered to be relatively immobile, should provide a reasonably accurate classification. The Nb/Y ratio is used as an indicator of alkalinity and the Zr/TiO_2 ratio serves as a differentiation index.

Metavolcanic rocks from the Mystic Lake-West Arm, Schist Lake area have also been silicified to varying degrees and, in the more extreme cases, this is clearly evident in the field (Reilly, personal communication). In general, samples chosen for geochemical analysis showed little or no obvious signs of silicification. Comparison of SiO_2 and immobile element distributions on the classification diagrams indicate; however, the alteration is more extensive than is apparent in the field.

Classification by immobile trace and minor elements (Figure 2) indicates that the vast majority of metavolcanics from this area are basaltic and sub-alkaline. Of the samples analysed to date, only two are more felsic than basalt and, in spite of the limited numbers of samples, this is considered to be generally representative of the distribution of metavolcanic rock types in the area (Reilly, personal communication).

The Jensen cation plots (Figures 3 and 4) further point to the dominantly basaltic nature of the metavolcanics in the two areas and also indicate that the majority are tholeiitic but with a lesser calc-alkaline component in the Douglas Lake-Phantom Lake area. This area also shows a tendency towards a bimodal distribution of compositions (Figure 3).

Metavolcanic rocks from the Douglas Lake-Phantom Lake area have low Ti contents characteristic of volcanic arc types (Figure 5). The majority of the metavolcanics are relatively unevolved, as would be expected from a suite of dominantly basaltic rocks. The suite as a whole shows a trend typical of arc sequences in that, with increasing Zr content, Ti concentrations show an initial increase but then drop off sharply in the more evolved (mainly intermediate to felsic) members of the suite.

Interpretation of the tectonic affinities of metavolcanic rocks from the Mystic Lake-West Arm, Schist Lake area in terms of Ti and Zr is more difficult. The distribution of data points (Figure 5) indicates that many of the basaltic rocks and the few evolved types show the low Ti con-

---

[1] Saskatchewan project A.114, from which this report derives, was funded in 1991 under the Canada-Saskatchewan Partnership Agreement on Mineral Development 1990-95.

[2] Department of Geology, University of Regina, Regina, Saskatchewan

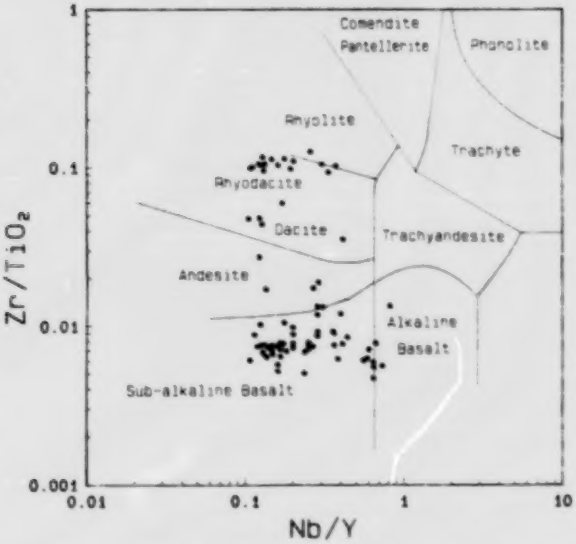


Figure 1 - Nb/Y vs. Zr/TiO$_2$ plot for metavolcanic rocks from the Douglas Lake-Phantom Lake area. Field boundaries are from Winchester and Floyd (1977).

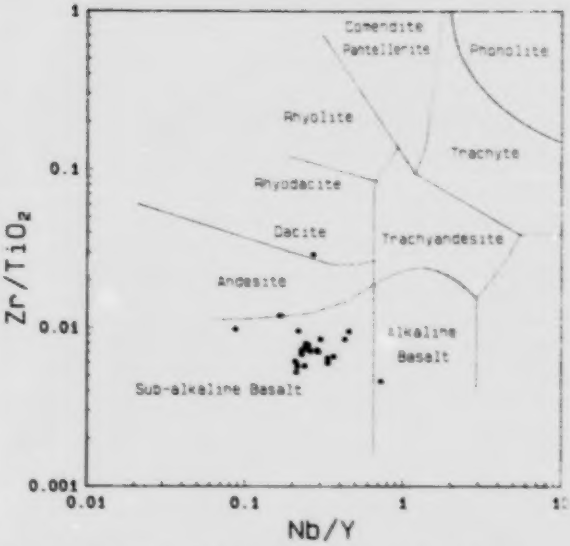


Figure 2 - Nb/Y vs. Zr/TiO$_2$ plot for metavolcanic rocks from the Mystic Lake-West Arm, Schist Lake area. Field boundaries are from Winchester and Floyd (1977).

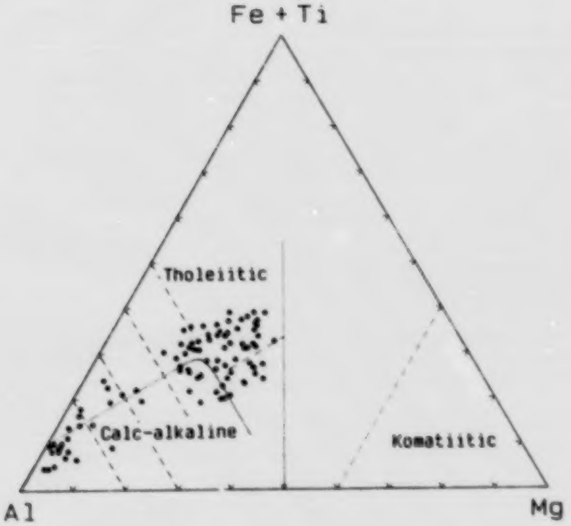


Figure 3 - Jensen cation plot for metavolcanic rocks from the Douglas Lake-Phantom Lake area. Field boundaries are from Jensen (1976).

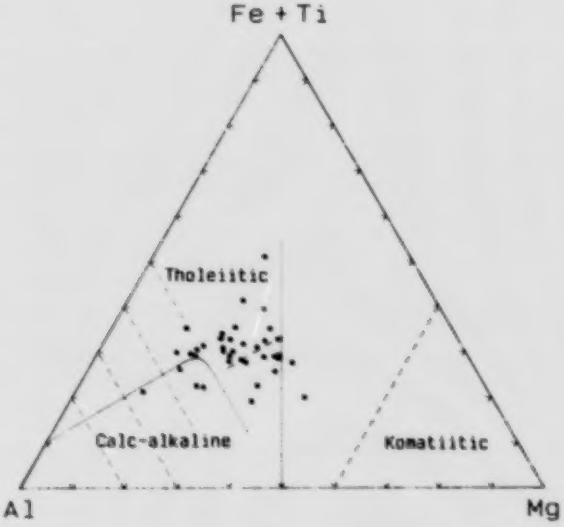


Figure 4 - Jensen cation plot for metavolcanic rocks from the Mystic Lake-West Arm, Schist Lake area. Field boundaries are from Jensen (1976).

tents typical of volcanic arc rocks. However, a significant number of the basaltic rocks have Ti contents (notably higher than many of the Douglas Lake-Phantom Lake tholeiitic basalts), that are characteristic of MORB and some within-plate basalts. The data are still too limited for any clear conclusions to explain this ambiguity. These compositional characteristics are, however, similar to the metabasalts from the Snake Rapids area along the west side of Amisk Lake (Watters and Ashton, this volume) and from the East Amisk area (Gaskarth and Parslow, 1987). It has been speculated that the tholeiitic basalts with relatively high Ti in these areas may have been emplaced in a transitional environment, possibly associated with back-arc spreading.

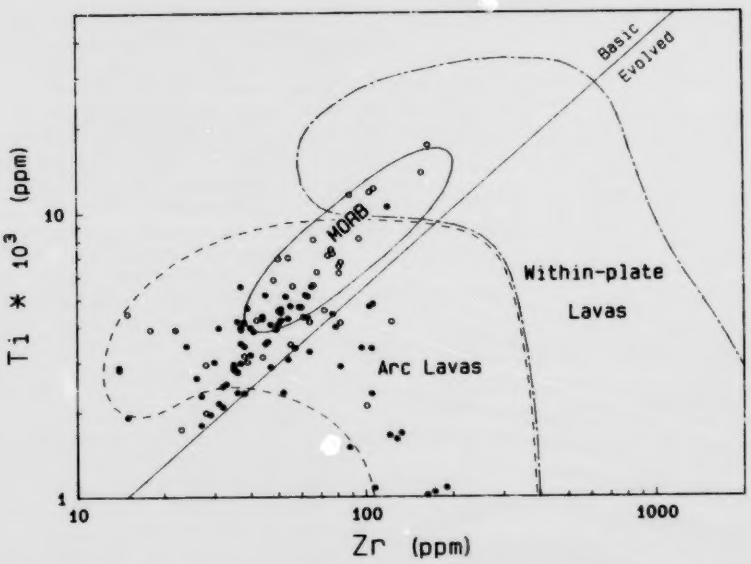


Figure 5 - Plot of Ti vs. Zr for metavolcanic rocks from the Douglas Lake-Phantom Lake area (filled circles) and the Mystic Lake-West Arm, Schist Lake area (open circles). Field boundaries are from Pearce and Norry (1979).

One of the main characteristics of island arc tholeiites is the low content of immobile incompatible elements, Ti and Y, for example. Watters (1990) has already shown that metavolcanic units from the Douglas Lake-Phantom Lake area have low Y contents for a given degree of fractionation (as indicated by the Cr content). Although many of the metabasalts from the Mystic Lake-West Arm, Schist Lake area show similar characteristics when plotted on a Y vs. Cr diagram, it is also apparent that the majority have significantly higher Y and Cr concentrations compared to the Douglas Lake-Phantom Lake metabasalts (Figure 6). Furthermore, many of the Mystic Lake-West Arm, Schist Lake metabasalts show Cr and Y contents that also typify both MORB and within-plate basalts, thereby providing a similarly ambiguous classification as did Ti and Zr.

Pearce (1982) showed that a convenient means of comparing and discriminating among basalts of different tectonic affinities is to plot their trace and minor element compositions as geochemical patterns, using typical mid-ocean ridge basalt (MORB) as a normalizing composition. The array of elements used for such patterns fall mainly into two groups: those which have a low ionic potential and are relatively mobile (Sr, K, Rb and Ba) and those with high ionic potential and relative immobility (Th, Ta, Nb, Ce, P, Zr, Hf, Sm, Ti, Y, Yb, Sc and Cr). The typical MORB (tholeiitic) pattern will be flat with abundance ratios close to unity. Within-plate basalts exhibit generally enriched patterns across much of the array, as do alkaline basalts and basalts from transitional environments. Patterns of volcanic arc basalts are characterized by a general enrichment of the incompatible elements of low ionic potential relative to tholeiitic MORB. Island arc tholeiites are depleted in the relatively immobile elements (high ionic potential) Th to Cr. Calc-alkaline basalts, although following these general patterns, also tend to be enriched in Th, Ce, P and Sm. Possible explanations, in terms of petrogenetic processes, for these differences in trace and minor element distribution patterns have been discussed by Pearce (1982).

MORB-normalized trace element patterns are plotted for selected samples of metabasalt from the Douglas Lake-Phantom Lake area (Figure 7) and from the Mystic Lake-West Arm, Schist Lake area (Figure 8). Although some of the elements have yet to be determined for the rocks of the present study, the data patterns provide a preliminary interpretation of the tectonic affinities of the Amisk Group metabasalts.

Metabasalts from the Douglas Lake-Phantom Lake area show a general pattern of moderate enrichment in the low ionic potential elements and moderate to strong depletion in the high ionic potential elements (Figure 7). Such patterns are similar to patterns for modern island arc tholeiites, whereas basalts that have relatively strong enrichment of the mobile elements represent calc-alkaline arc basalts.

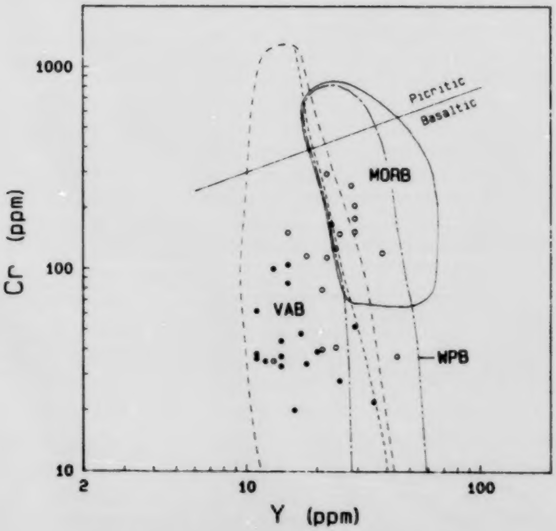


Figure 6 - Plot of Cr vs. Y for metabasaltic rocks from the Douglas Lake-Phantom Lake area (filled circles) and the Mystic Lake-West Arm, Schist Lake area (open circles). VAB = volcanic arc basalt; MORB = mid-ocean ridge basalt; WPB = within-plate basalt. Field boundaries are from Pearce (1982).

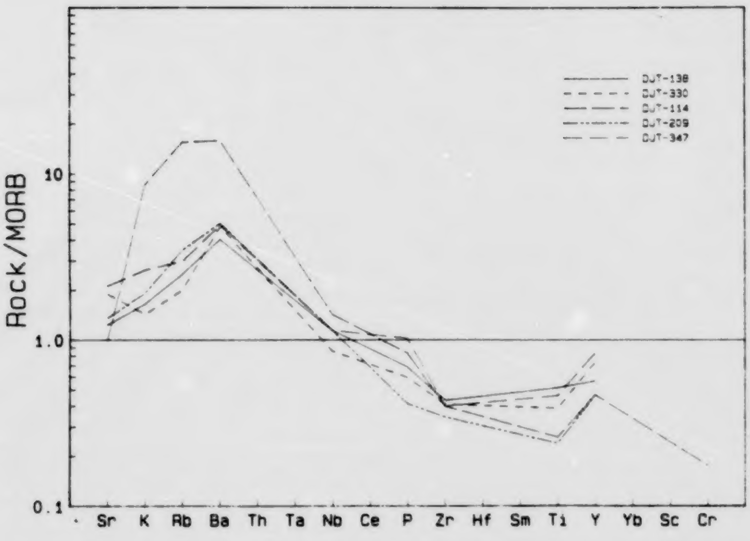


Figure 7 - Trace and minor element distribution patterns (MORB normalized) for some typical metabasalts from the Douglas Lake-Phantom Lake area. Normalizing values are from Pearce (1982).

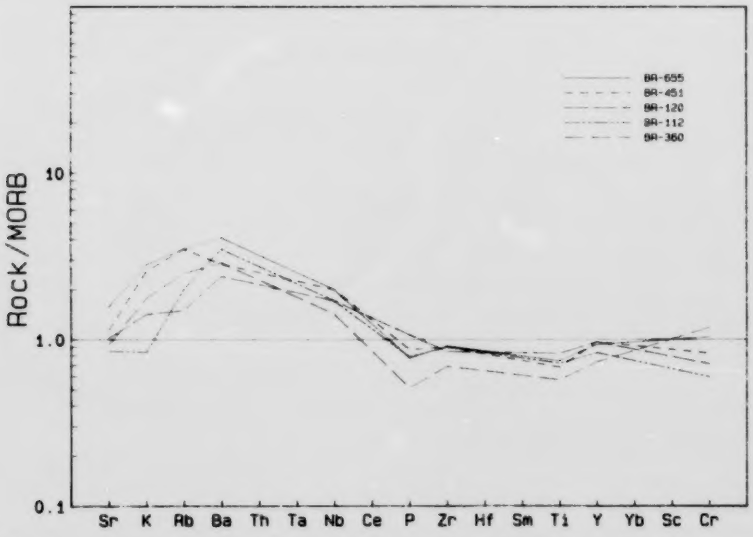


Figure 8 - Trace and minor element distribution patterns (MORB normalized) for some typical metabasalts from the Mystic Lake-West Arm, Schist Lake area. Normalizing values are from Pearce (1982).

The geochemical patterns, displayed by metabasalts from the Mystic Lake-West Arm, Schist Lake area (Figure 8) are, in general, somewhat different in that they show relatively low enrichment of the mobile elements and a small degree of depletion of the immobile elements. Although such patterns can still be considered typical of some island arc tholeiites, they are also similar to patterns displayed by MORB from some transitional environments.

These geochemical patterns once again point to differences between some of the metabasalts in the Mystic Lake-West Arm, Schist Lake area and those of the Douglas Lake-Phantom Lake area suggesting, perhaps, that the tectonic affinities of the metabasalts from the former area are more complex than previously thought. More detailed geochemical data is needed in order to resolve the present ambiguities in interpretation of the geotectonic setting.

3. Boot Lake-Phantom Lake Intrusion

The Boot Lake-Phantom Lake Intrusion, as described by Galley and Franklin (1987, 1989) and Thomas (1989), is an intrusive complex comprising five main phases, ranging from early mafic to successively more felsic phases. The phases are: Phase 1 - mainly of gabbroic to dioritic rocks; Phase 2 - hornblende-biotite porphyritic quartz monzodiorite and minor biotite porphyritic syenite; Phase 3 - tonalite; Phase 4 - plagioclase-porphyritic granodiorite; Phase 5 - microcline-porphyritic granodiorite.

The complex was mapped by Byers *et al* (1965) and Pearson (1984) as two separate bodies, the Boot Lake granodiorite (mainly Phases 2, 3 and 4) and the Phantom Lake porphyritic granite (Phase 5). Byers *et al* (1965) suggested that the Boot Lake granodiorite had been intruded during an earlier deformational event than the Phantom Lake body. Galley and Franklin (1987) noted gradational contacts between the microcline-porphyritic granodiorite and the earlier granodiorite and, on that evidence, suggested a relatively short time span between emplacement of the two intrusions.

The recent remapping of the Boot Lake-Phantom Lake Intrusion by Thomas (1989) indicates that, whereas the earliest phases are heterogeneous with largely gradational to diffuse contact relationships, the porphyritic granodiorite comprising Phase 5 intrudes the earlier phases sharply and appears to be a distinctive unit. Thomas (1991, personal communication) has suggested that, compared with the Phases 1, 2, 3 and 4, the Phantom Lake granodiorite may represent a petrologically distinct late intrusion.

In terms of $Na_2O + K_2O$, FeO and MgO, the intrusive rocks are dominantly calc-alkaline and form a continuous compositional spectrum from gabbroic to granitic (Figure 9). The felsic end of the spectrum on this diagram is represented by the porphyritic granodiorite of Phase 5, which does not appear as a distinct unit, although the compositional overlap with other phases is small.

On a plot of $Na_2O + K_2O$ versus SiO_2 (Figure 10), the early phases define a compositional trend of increasing alkali content with increasing SiO_2 that does not include the porphyritic granodiorite rocks of Phase 5. The latter are distinctly more enriched in the alkali elements than are the earlier phases, even for those samples which have higher SiO_2 than the porphyritic granodiorite. All phases are sub-alkaline. The relative compositional homogeneity of Phase 5 is evident from the narrow range of SiO_2 values (64.22 to 67.45 percent) which is consistent with granodiorite classification.

The contents of certain trace elements, Ba and Sr in particular, illustrate even more clearly the distinctiveness of the late porphyritic granodiorite phase. Plots of SiO_2 against Sr and CaO against Sr (Figures 11 and 12) indicate the relatively very high concentrations of Sr in the granodiorite (1280 to 1680 ppm) compared with those in the early phases (127 to 609 ppm). The trend in the early phases is one of decreasing Sr with increasing silica content and increasing Sr with increasing calcium concentration. Such trends are expected, resulting from the generation of a series of magmas by varying degrees of partial melting of a common source rock and/or subsequent differentiation. They support the suggestion of co-magmatism for the early phases of the intrusive complex. The corresponding increase in Sr with increasing Ca indicates that the trace element has substituted for the major element primarily in plagioclase feldspar.

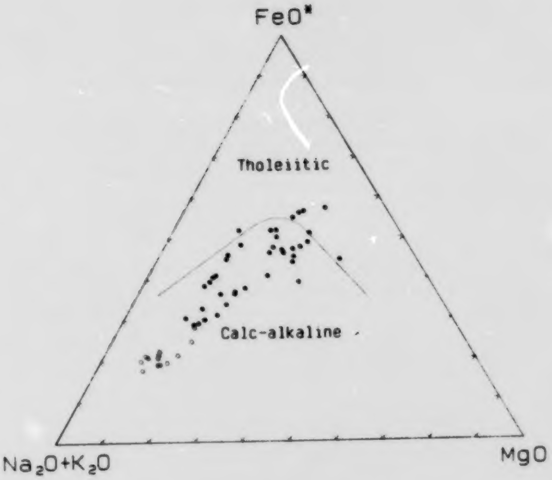


Figure 9 - AFM ($Na_2O + K_2O$-FeO-MgO) plot for rocks from the Boot Lake-Phantom Lake Intrusion. Field boundaries are from Irvine and Barager (1971). Filled circles = dominantly gabbroic to granodioritic early phases; open circles = Phantom Lake porphyritic granodiorite (Phase 5).*

By contrast, Phase 5 granodiorite shows no obvious compositional coherence between CaO and Sr, although the range of composition may be too small for positive covariation to be apparent. It is clear from Figure 12 that Phase 5 granodiorite contains very high concentrations of Sr in relation to CaO content and, therefore, does not form part of the same magma series as the earlier phases. The magma giving rise to the Phase 5 body must have been generated by partial melting of different source rocks which were also, presumably, enriched in Sr.

The element Ba also shows clearly that Phase 5 granodiorite is quite distinct from the other phases of the intrusive complex. Relationships between Ba and silica (not shown here) are similar to those between Sr and silica. A plot of Ba against the immobile element Zr (Figure 13) indicates that the early phases form a well-defined trend of increasing Ba content (181 to 1033 ppm) with increasing Zr content (28 to 183 ppm) (correlating

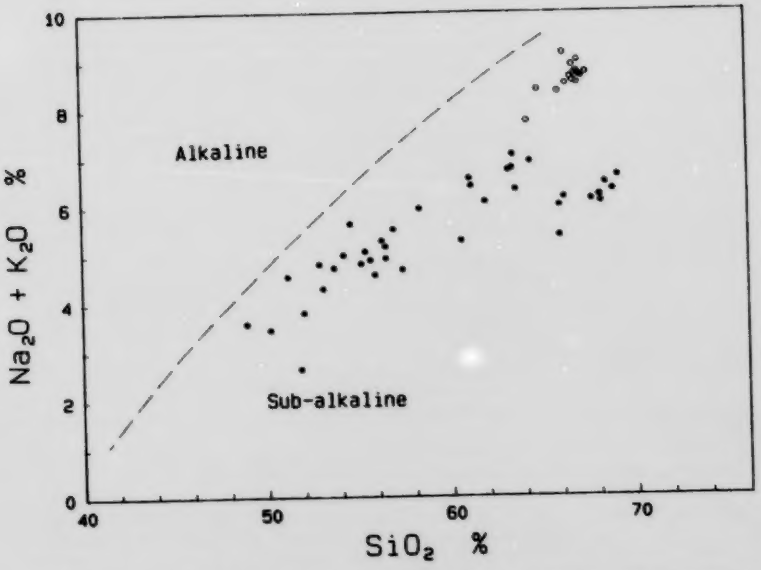


Figure 10 - SiO_2 vs. $Na_2O + K_2O$ plot for rocks from the Boot Lake-Phantom Lake Intrusion. Field boundaries are from Irvine and Barager (1971). Symbols as for Figure 9.

with an increase in silica). Phase 5 rocks clearly do not conform to this compositional trend and have significantly higher Ba concentrations (1361 to 1900 ppm) as well as generally high levels of Zr (123 to 191 ppm).

Pearce *et al.* (1984) have shown that it is possible to determine the tectonic setting of emplacement of granitic bodies on the basis of their trace element characteristics. Applying these criteria to the Boot Lake-Phantom Lake Intrusion (Figure 14) indicates that the magmas were emplaced in a volcanic arc environment which would be consistent with their intrusion into the Amisk Group volcanic succession.

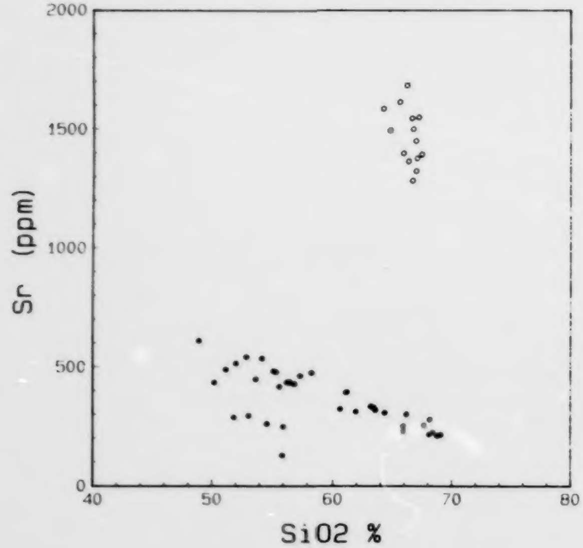


Figure 11 - SiO_2 vs. Sr plot for rocks from the Boot Lake-Phantom Lake Intrusion. Symbols as for Figure 9.

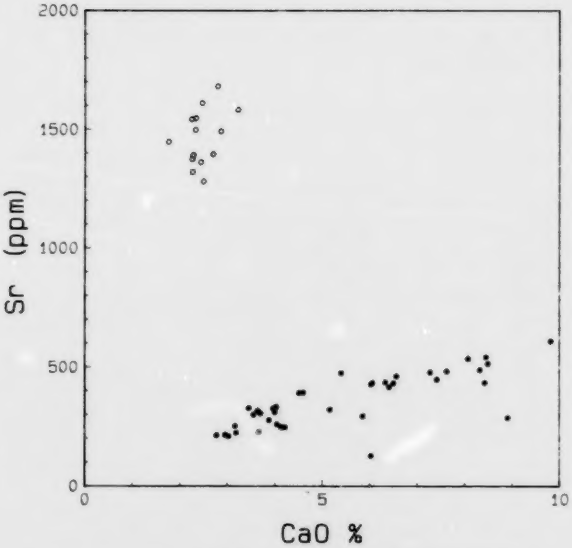


Figure 12 - CaO vs. Sr plot for rocks from the Boot Lake-Phantom Lake Intrusion. Symbols as for Figure 9.

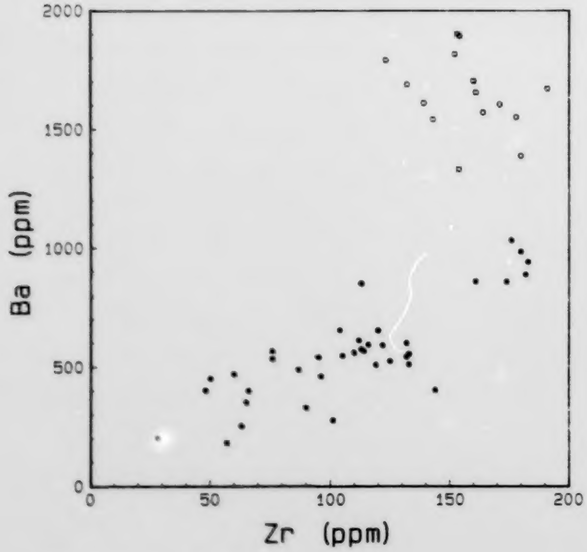


Figure 13 - Zr vs. Ba plot for rocks from the Boot Lake-Phantom Lake Intrusion. Symbols as for Figure 9.

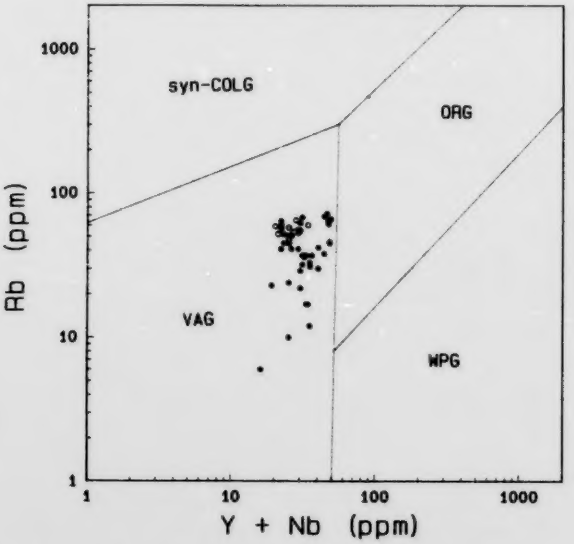


Figure 14 - Rb vs. Y+Nb discriminant plot for rocks from the Boot Lake-Phantom Lake Intrusion. Field boundaries are from Pearce et al. (1984). VAG=volcanic arc granites; WPG=within-plate granite; ORG=ocean ridge granite; syn-COLG=syn-collision granite. Symbols as for Figure 9.

4. Conclusions

1. Metavolcanic rocks from the Douglas Lake-Phantom Lake area are dominantly basaltic, with minor intermediate and rhyolitic types. The suite is mainly tholeiitic, although lesser calc-alkaline types are also present, and shows volcanic arc affinities.

2. Metavolcanic sequences in the Mystic Lake-West Arm, Schist Lake area comprise basaltic rocks almost exclusively and, whereas most can be classified as island arc tholeiites, a significant proportion of the suite show trace element characteristics that are also typical of tholeiitic MORB and within-plate basalts. Concentrations of the immobile elements, such as Ti, Y and Cr are notably higher in these latter rocks when compared to the rest of the suite and with metabasalts from the Douglas Lake-Phantom Lake area.

3. The early phases of the Boot Lake-Phantom Lake Intrusion display a coherent and continuous compositional trend from mafic to felsic and would appear to form a co-magmatic series.

4. The latest phase of the intrusion, Phase 5, comprising porphyritic granodiorite has certain compositional characteristics which are distinct from the earlier phases. These rocks are marked by a fairly restricted compositional range, relatively high total alkalis and relatively high concentrations of some trace elements, notably Sr and Ba. These compositional differences are significant and suggest the involvement of different source rocks and/or magmatic processes in the evolution of the Phantom Lake granodiorite.

5. The compositional distinctiveness of the porphyritic granodiorite, and the consequent implications regarding magma genesis, suggest that the latest phase (Phase 5) of this intrusive complex should be regarded as a separate intrusion and not part of a co-magmatic series with the early phases.

6. The Boot Lake-Phantom Lake Intrusion can be classified as a "volcanic arc granite".

5. References

Ashton, K.E. (1990): Geology of the Snake Rapids Area, Flin Flon Domain (parts of NTS 63L-9 and -10); *in* Summary of Investigations 1990, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 90-4, p4-12.

Byers, A.R., Kirkland, S.J.T. and Pearson, W.J. (1965): Geology and mineral deposits of the Flin Flon area, Saskatchewan; Sask. Dep. Miner. Resour., Rep. 62, 95p.

Galley, A.G. and Franklin, J.M. (1987): Geological setting of gold, copper, tungsten and molybdenum occurrences in the Phantom Lake region; *in* Summary of Investigations 1987, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 87-4, p115-124.

__________ (1989): Setting of gold in the Phantom Lake region, Saskatchewan; *in* Investigations by the Geological Survey of Canada in Manitoba and Saskatchewan during the 1984-1989 Mineral Development Agreement, Geol. Surv. Can., Open File Rep. 2133, p21-33.

Gaskarth, J.W. and Parslow, G.R. (1987): Proterozoic volcanism in the Flin Flon greenstone belt, east-central Saskatchewan, Canada; *in* Pharaoh, T.C., Beckinsale, R.D. and Rickard, D. (eds.), Geochemistry and Mineralization of Proterozoic Volcanic Suites, Geol. Soc., Spec. Publ. No. 33, p183-200.

Irvine, T.N. and Barager, W.R. (1971): A guide to the chemical classification of the common volcanic rocks; Can. J. Earth Sci., v8, p523-548.

Jensen, L.S. (1976): A new cation plot for classifying subalkalic volcanic rocks; Ont. Geol. Surv., Misc. Pap. 66.

Pearce. J.A. (1982): Trace element characteristics of lavas from destructive plate boundaries; *in* Thorpe, R.S. (ed.), Andesites, J.Wiley and Sons, New York, p525-547.

Pearce, J.A., Harris, N.B.W. and Tindle, A.G. (1984): Trace element discrimination diagrams for the tectonic interpretation of granitic rocks; J. Petrol., v25, p956-983.

Pearce, J.A. and Norry, M.A. (1979): Petrogenetic implications of Ti, Zr, Y and Nb variations in volcanic rocks; Contrib. Mineral. Petrol., v69, p33-47.

Pearson, J.G. (1984): Gold metallogenic studies: the Rio deposit; *in* Summary of Investigations 1984, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 84-4, p123-126.

Reilly, B.A. (1990): Bedrock geological mapping, Mystic Lake-West Arm, Schist Lake Area (Part of NTS 63K-12); *in* Summary of Investigations 1990, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 90-4, p25-35.

Thomas, D.J. (1989): Geology of the Douglas Lake-Phantom Lake area (part of NTS 63K-12 and -13); *in* Summary of Investigations 1989, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 89-4, p44-54.

__________ (1990): New perspectives on the Amisk Group and regional metallogeny, Douglas Lake-Phantom Lake area, northern Saskatchewan; *in* Summary of Investigations 1990, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 90-4, p13-20.

Watters, B.R. (1989): Lithogeochemical studies in the Flin Flon-Amisk Lake area and Glennie Domain; *in* Summary of Investigations 1989, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 89-4, p55-57.

__________ (1990): Lithogeochemical studies in the Flin Flon-Amisk Lake area; *in* Summary of Investigations 1990, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 90-4, p44-47.

Winchester, J.A. and Floyd, P.A. (1977): Geochemical discrimination of different magma series and their differentiation products using immobile elements; Chem. Geol., v20, p325-343.

The NatMap Shield Margin Project and Project Seagull in Saskatchewan: An Integrated Geoscience Program Incorporating GIS[1]

R. Macdonald, S. Lucas[2], D.J. Thomas and B.M. Czornobay

Macdonald, R., Lucas, S., Thomas, D.J. and Czornobay, B.M. (1991): The NatMap Shield Margin Project and Project Seagull in Saskatchewan: an integrated geoscience program incorporating GIS; *in* Summary of Investigations 1991, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 91-4.

The Flin Flon-Snow Lake-Hanson Lake region of the Precambrian Shield and the area lying immediately south of the Shield in Saskatchewan and Manitoba, where the Phanerozoic cover is relatively thin, is of significant interest to the mineral exploration industry, as it is known to contain Cu-Ni-Zn massive sulphide and gold deposits and has considerable potential for further discoveries.

In 1989 the Saskatchewan Geological Survey (SGS) initiated a program to remap the bedrock geology of the Flin Flon area in Saskatchewan at scales of 1:12 500 to 1:20 000. In 1990 this program was incorporated into the Canada-Saskatchewan Partnership Agreement on Mineral Development 1990-95 as Project Seagull. In 1990-91 the Geological Survey of Canada (GSC) initiated the Shield Margin Project of the national Mapping Program (NATMAP) to provide for geoscience studies in the region. In Spring 1991 a collaborative agreement was established to coordinate the GSC NATMAP Shield Margin geoscience program with the SGS Project Seagull in the same area. Similar agreements were reached between the GSC and the Manitoba Geological Services Branch (MGSB) and the Universities of Regina, Saskatchewan, Manitoba and New Brunswick. A major transect of the Lithoprobe program in the Trans-Hudson Orogen, which commenced this summer, also covers part of the Shield Margin project area.

An essential product of this joint initiative is to incorporate geoscience data from this collaboration into a geographical information system (GIS). This GIS is planned to comprise a comprehensive digital geoscience database housing all observational, compilatory and interpretive information, including bedrock geological, geochronological, isotopic, structural, drillcore and high resolution geophysical data. A digital database working group with representatives from major project participants has already commenced work on design and implementation. The GIS database is planned to be made available for use in the Public domain through published hard copy products such as maps and reports, as well as in direct digital formats. The first product will be a pilot 1:50 000 coloured geological map of the East Amisk area demonstrating current GSC and SGS computerized mapping for display at the 1991 SGS Open House and at the GSC Minerals Colloquium in January 1992. To facilitate this, SGS geologists commenced collecting field data this summer using FIELDLOG, a software package provided by the Ontario Geological Survey which enables the production of geological maps linked to a geographical database. Data collected during previous seasons has also been compiled into this database.

---

(1) Saskatchewan Project A.11 (Project Seagull), from which this report derives, was funded in 1991 under the Canada-Saskatchewan Partnership Agreement on Mineral Development 1990-95.

(2) Continental Geoscience Division, Geological Survey of Canada, Ottawa, Ontario

Revision Bedrock Geological Mapping, Wildnest-Attitti Lakes Area (Parts of NTS 63M-1 and -2)[1]

K.E. Ashton and A.D. Leclair[2]

Ashton, K.E. and Leclair, A.D. (1991): Revision bedrock geological mapping, Wildnest-Attitti Lakes area (parts of NTS 63M-1 and -2); in Summary of Investigations 1991, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 91-4.

This paper reports on the first part of an intended three-year, 1:20 000 scale mapping project aimed at updating and reinterpreting the regional geology along a 55 x 12 km transect between Wildnest Lake and the Tabbernor Fault zone (Figure 1). The summer of 1991 was spent in the easternmost segment of the transect in the Wildnest-Attitti Lakes area. The map area is centred about 45 km northwest of Flin Flon, Manitoba and 40 km southeast of Pelican Narrows. It is accessible by float-equipped aircraft or by canoe routes from either centre. Field work was carried out by a five-person crew and consisted of hip chain and compass traverses and shoreline mapping. Topographic relief ranges over 50 m with supracrustal rocks generally occupying the low areas and granitoid rocks dominating the ridges.

Early geological reconnaissance, carried out by McInnes (1914), Satterly (1932) and Taylor (1958), was followed by 1:63,360 scale geological mapping by Pyke (1961, 1966). The Wildnest-Tabbernor Transect represents the western continuation of a remapping program carried out as part of the Canada-Saskatchewan Mineral Development Agreement 1984-1989 (Ashton and Wheatley, 1986; Ashton *et al.*, 1986, 1987; Ashton, 1987, 1989a, 1990a, in prep.).

1. General Geology

The Wildnest-Attitti Lakes area is located in the central part of the Attitti Block (Macdonald, 1981), a subdomain of the Kisseynew Domain (Figure 1). The Attitti Block is separated from the Hanson Lake Block to the west by the Sturgeon-weir Shear Zone (MacQuarrie, 1979; Ashton *et al.*, 1987; Ashton and Wilcox, 1988), and from the rest of the Kisseynew Domain to the east by a gradual change from dominantly hornblende- to biotite-rich gneisses (Macdonald, 1981).

The Wildnest-Attitti Lakes area is underlain by an upper amphibolite facies metamorphic suite of supracrustal rocks and intruded granitoid plutons[3]. Primary features have generally been obliterated by the combined effects of metamorphism and deformation. A complex deformational history is suggested by multiple generations of folding and variably developed shearing.

Pyke (1961, 1966) interpreted the supracrustal rocks as predominantly sedimentary with only a minor mafic component of possible volcanic origin. In a 1:250 000 scale compilation of the Pelican Narrows and Amisk Lake areas, Macdonald (1981) treated the supracrustal rocks of the Attitti Block as largely volcanogenic. Based on 1:50 000 scale mapping in the Granite Lake area, Ashton *et al.* (1987) confirmed a previous suggestion by Byers and Dahlstrom (1954) that supracrustal rocks of the Flin Flon Domain could be traced continuously along strike into the Attitti Block of the Kisseynew Domain. This allowed for broad stratigraphic correlation of the supracrustal rocks and extended usage of the Amisk-Missi (Bruce, 1918; Wright, 1933) stratigraphic classification (Ashton, 1989a).

This summer's findings are consistent with the interpretation of the Attitti Block as a high-grade equivalent of the Flin Flon Domain. Rare relict primary features, rock compositions and associations, the near-ubiquitous presence of disseminated pyrrhotite, and scattered occurrences of rocks affected by Mg-metasomatism all suggest that supracrustal rocks of the Attitti Block were derived from predominantly volcanic precursors.

The boundary between the Attitti Block and Kisseynew Domain is interpreted as a facies change from dominantly volcanic to dominantly sedimentary rocks, respectively, although it is likely complicated by structural telescoping prior to the peak of metamorphism. The nature of the facies change, together with the inferred structural complications, make it difficult to define a meaningful boundary.

Pyke (1961, 1966) originally assigned metamorphic names to the supracrustal rock units due to the difficulty in determining precursors. During this project, an attempt was made to "see through" the effects of metamorphism and assign primary lithological rock names. Although it was generally not possible to distinguish between flows and volcaniclastic rocks, detailed mapping has allowed further refinement and subdivision of both the supracrustal and granitoid rocks. Classification and nomenclature of plutonic rocks follows Streckeisen (1976).

---

(1) Saskatchewan Project A.121, from which this report derives, was funded in 1991 under the Canada-Saskatchewan Partnership Agreement on Mineral Development 1990-95.

(2) Geological Survey of Canada, 601 Booth St., Ottawa, Ontario K1A 0E8.

(3) Although the prefix "meta-" has not been used, it is understood that all rock types in the area have undergone high-grade metamorphism with the exception of the undeformed medium-grained to pegmatitic granitoid rocks and Paleozoic rocks.

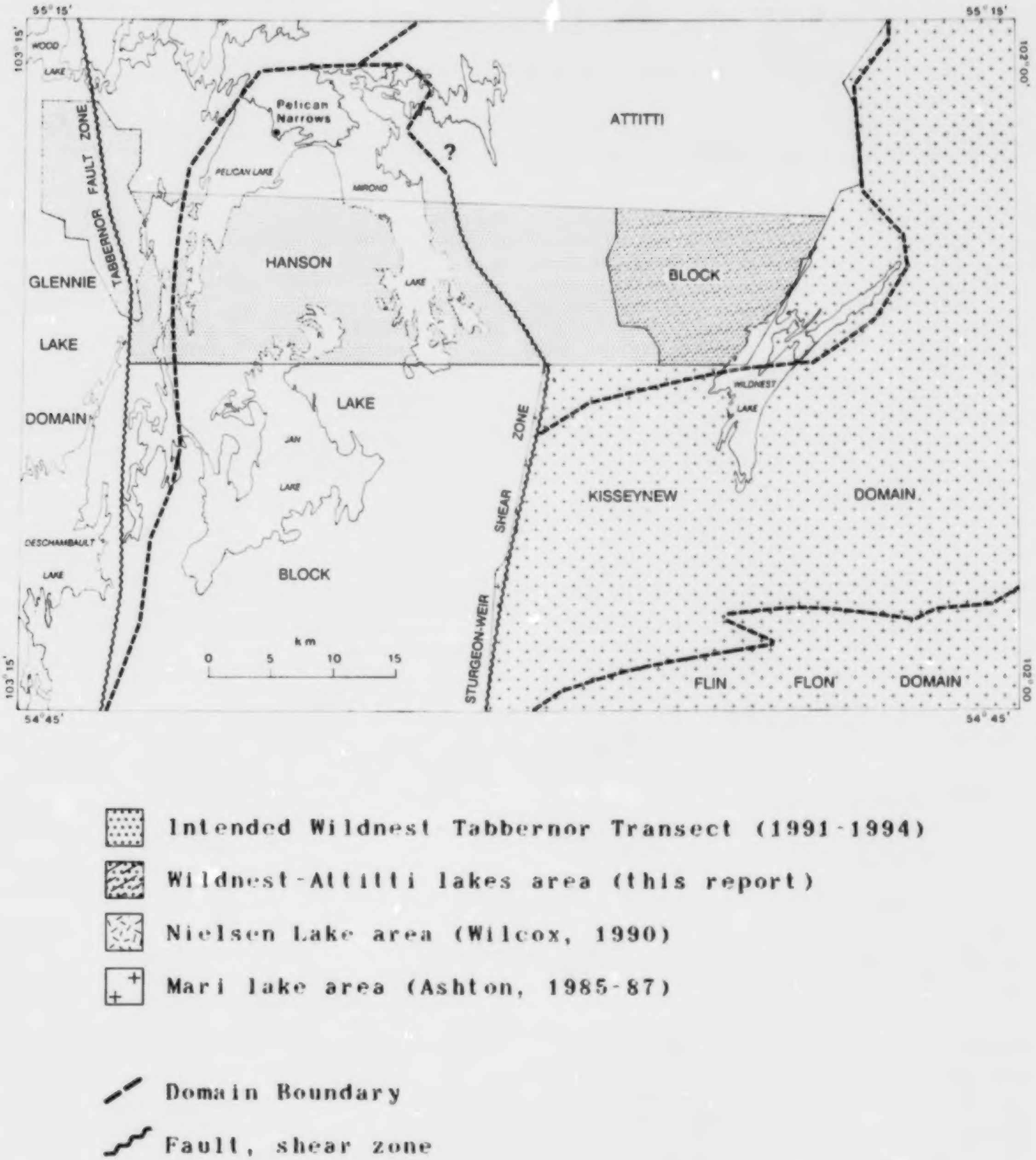

55° 15'
103° 15'
WOOD LAKE
TABBERNOR FAULT ZONE
Pelican Narrows
PELICAN LAKE
MIROND
ATTITTI
?
HANSON
LAKE
BLOCK
GLENNIE
LAKE
DOMAIN
LAKE
JAN
LAKE
BLOCK
WILDNEST
LAKE
STURGEON-WEIR SHEAR ZONE
KISSEYNEW
DOMAIN
DESCHAMBAULT LAKE
km
0
5
10
15
FLIN
FLON
DOMAIN
54° 45'
102° 00'
Intended Wildnest Tabbernor Transect (1991-1994)
Wildnest-Attitti lakes area (this report)
Nielsen Lake area (Wilcox, 1990)
Mari lake area (Ashton, 1985-87)
Domain Boundary
Fault, shear zone

Figure 1 - Location of recent and proposed work (modified from domain map of Macdonald, 1981).

2. Unit Descriptions

a) Amisk Group Volcanic and Associated Rocks

A "metamorphic colour index" based on the percentage of mafic minerals in a rock was used in the field to name rocks of the volcanic suite. An index of 40 was used to distinguish mafic from intermediate rocks and 20 was used for the intermediate/felsic boundary.

Intermediate to Mafic Volcanic Suite

Dark green to black mafic rocks (Am) form units up to several hundred metres in width over much of the map area . Most rocks are fine- to medium-grained, homogeneous, and weakly foliated. They consist of approximately equal proportions of hornblende and plagioclase with minor clinopyroxene. Up to 15 percent red garnet is present at some outcrops and locally defines metre-scale layers. Magnetite-rich layers are spatially associated with the garnetiferous rocks at some occurrences. Rocks of intermediate composition (Ai) are commonly interlayered with the mafic rocks. They contain variable proportions of hornblende, biotite, quartz and plagioclase with local garnet and cummingtonite.

Pale green medium- to coarse-grained lenses containing 10 to 50 percent clinopyroxene, 0 to 20 percent garnet, 0 to 10 percent carbonate, 30 to 60 percent plagioclase and minor quartz comprise 0 to 30 percent of the intermediate to mafic rocks. They most commonly occur as centimetre-scale lenses making up about 10 percent of the rock, but locally form layers which are continuous over several metres imparting a well layered appearance. Sulphides tend to be concentrated in these clinopyroxene-bearing lenses and layers.

Calc-silicate layers and lenses distinguished by the presence of abundant carbonate are also associated with many intermediate to mafic rocks (see accompanying map), but are rarely of sufficient extent to appear as a separate unit on a 1:20 000 scale map. They contain variable proportions of clinopyroxene, Ca-rich amphibole, sphene, carbonate, quartz, plagioclase and local garnet, scapolite, graphite and pyrrhotite. These calcareous rocks are thought to be derived by pre-metamorphic carbonatization of rocks of the volcanic suite.

Homogeneous fine- to medium-grained mafic rocks along the north side of Rushton Lake (Figure 2) contain rare hornblendite layers less than 1 m thick. Their homogeneous nature and ultramafic association suggest that they are intrusive in part.

Pyke (1961) suggested that the more homogeneous intermediate to mafic rocks were volcanic in origin and that the well-layered varieties were derived from sedimentary precursors. Volcaniclastic rocks are represented, but many of the layered rocks consist of thin layers of hornblende and plagioclase alternating with the pale green clinopyroxene-plagioclase layers described above. It is unlikely that the clinopyroxene-plagioclase layers reflect a primary igneous or sedimentary composition. They are probably the high-grade equivalents of altered material and may represent the pervasive epidote-rich alteration which characterizes many mafic volcanic rocks of the Flin Flon Domain. Their layered nature could result from transposition during deformation.

The intermediate to mafic volcanic suite is thought to include the high-grade equivalents of volcanic and volcaniclastic rocks of andesitic to basaltic composition. Some of the more homogeneous rocks are probably derived from diabases and gabbros.

Felsic to Intermediate Volcanic Suite

Pyke's (1961) biotite gneiss and granulite unit, which underlies 10 to 15 percent of the area, has been broadly interpreted in this work as a felsic (Af) to intermediate (Ai) volcanic suite. Most occurrences are spatially associated with more mafic rocks (Am) and locally mark the contact between such rocks and wackes.

The felsic rocks (Af) contain 0 to 20 percent combined biotite and hornblende with local garnet and clinopyroxene. Minor pyrrhotite is nearly ubiquitous and the unit also contains numerous small gossans and associated rocks indicative of pre-metamorphic Fe-Mg metasomatism (see accompanying map). Quartz and feldspar phenocrysts are present at several localities providing evidence of the igneous nature of these rocks.

Intermediate compositions (Ai) include both relatively homogeneous rocks of probable dacitic origin and well-layered rocks likely derived from volcaniclastic precursors. They typically contain more hornblende than the felsic rocks with local garnet and cummingtonite.

The large body in the File Bay area (Figure 2) is more homogeneous than most occurrences and is distinguished by the presence of a clinopyroxene-bearing leucosome. The main rock type is light grey and contains 5 to 15 percent biotite with or without minor hornblende. Its homogeneous character suggests an intrusive, sub-volcanic origin for most of the body, although it is more heterogeneous towards its margins where it is interlayered with intermediate to mafic rocks and iron formation.

Like their more mafic counterparts, rocks of the felsic to intermediate volcanic suite locally grade into calc-silicate rocks with the appearance of clinopyroxene, sphene and carbonate. Calc-silicate units range in thickness from a few centimetres up to a few tens of metres, but they could not be traced into a continuous map-scale unit. This mode of occurrence coupled with their appearance within rocks ranging from felsic to mafic composition suggests that they are the metamorphosed equivalents of carbonatized volcanic and volcaniclastic rocks.

Pyke (1961) interpreted the rocks of his biotite gneiss and granulite unit as paragneisses based on their composition. They have been reinterpreted during this work as felsic to intermediate volcanic, volcaniclastic and subvolcanic rocks based on the rare presence of quartz and feldspar phenocrysts, their spatial association with mafic volcanic rocks, the presence of disseminated sul-

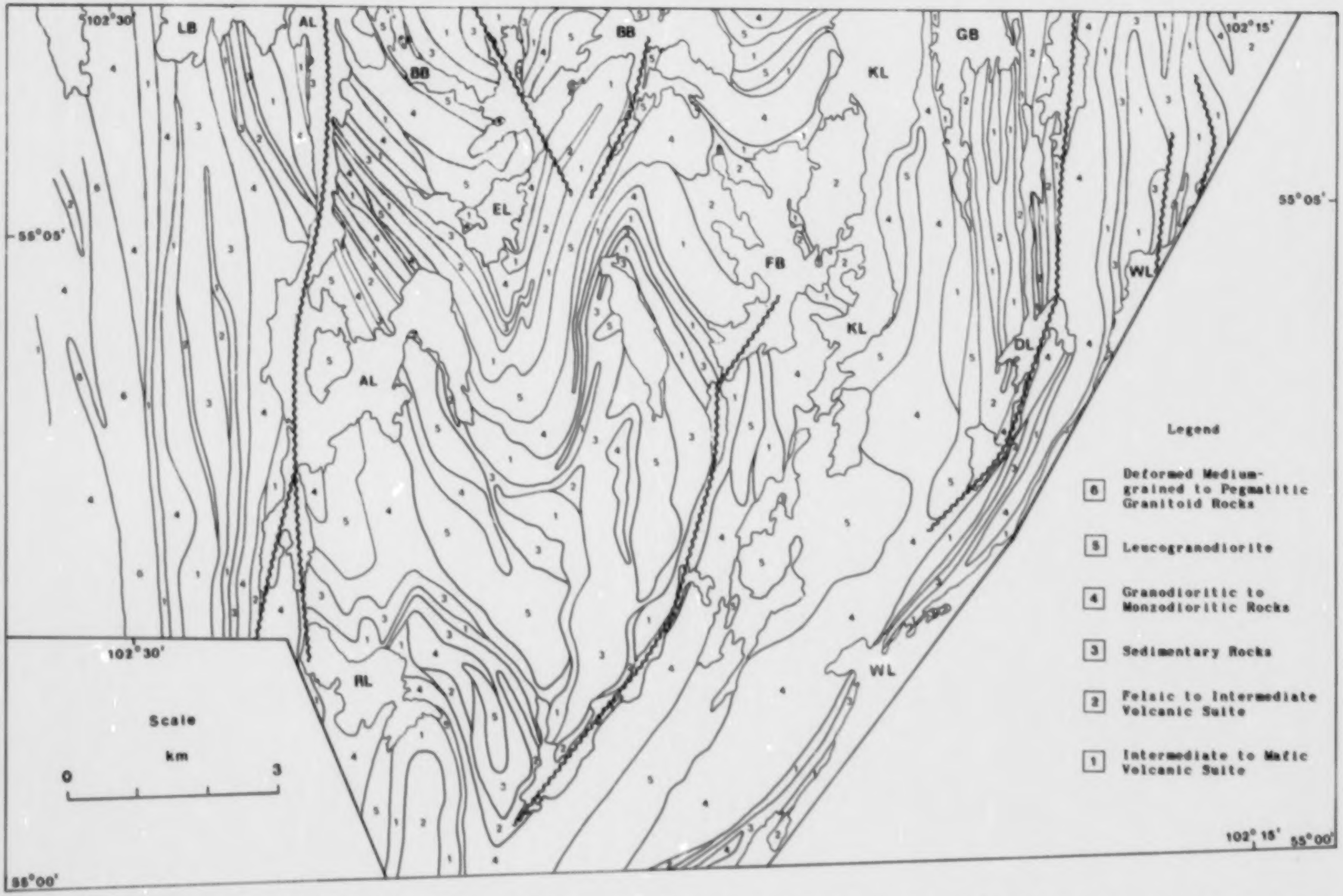

102°30'
LB
AL
BB
BB
EL
AL
RL
FB
KL
KL
GB
DL
WL
WL
102°15'
55°05'
55°05'
102°30'
102°15'
55°00'
55°00'
Scale
km
0
3
Legend
6 Deformed Medium-grained to Pegmatitic Granitoid Rocks
5 Leucogranodiorite
4 Granodioritic to Monzodioritic Rocks
3 Sedimentary Rocks
2 Felsic to Intermediate Volcanic Suite
1 Intermediate to Mafic Volcanic Suite

Figure 2 - Simplified geological map of the Wildnest-Attitti Lakes area (WL-Wildnest Lake, DL-Dougherty Lake, KL-Kakinagimak Lake, GB-Gifford Bay, FB-File Bay, EL-Ewen Lake, AL-Attitti Lake, BB-Bentz Bay, LB-Lowe Bay, RL-Rushton Lake).

phides and associated alteration, and their compositional similarity to lower-grade volcanic rocks. Most of the felsic rocks are dacitic in composition but rare occurrences containing less than 10 percent ferromagnesian minerals are probably rhyolitic.

A white-weathering unit of quartzofeldspathic biotite±garnet±sillimanite gneiss (Ar) was recognized in the Gifford Bay area of Kakinagimak Lake (see accompanying map). It has an irregular outcrop pattern due to interlayering over tens to hundreds of metres with rocks of the felsic to intermediate and intermediate to mafic volcanic suites. Contacts are generally sharp.

Most rocks are grey-white and contain about 10 percent biotite, 30 to 70 percent quartz and 30 to 70 percent feldspar, but more fissile layers containing up to 20 percent sillimanite with associated garnet and biotite are intercalated. The sillimanite-bearing rocks contain a medium- to coarse-grained, pink leucosome, possibly indicating the presence of K-feldspar and sillimanite due to the breakdown of muscovite. Layers of felsic to intermediate hornblende-bearing and garnet-biotite rocks less than a metre thick are intercalated with the sillimanite-bearing quartzofeldspathic rocks and display sharp contacts.

The irregular outcrop pattern, lithological association and extreme quartz- and sillimanite-rich composition of these rocks suggest that they were not derived from sedimentary precursors. They may have been derived in part from rhyolitic rocks, but they are thought to have been affected by pre-metamorphic silicification and aluminous alteration.

b) Amisk Group Sedimentary Rocks

Chemical sedimentary rocks are intercalated with rocks of the felsic to intermediate volcanic suite in several forms:

1) Sulphide-facies iron formation (Ans) is found in and along the contacts of the intermediate to mafic volcanic suite (see accompanying map). It occurs as thin layers containing several percent pyrrhotite and trace chalcopyrite in a fine-grained, sugary quartz-rich matrix. It is generally marked by a gossan and is similar to sulphide-facies iron formations found elsewhere in the Kisseynew Domain (Ashton, 1989b).

2) On the western shore of File Bay (see accompanying map), a thick unit of recrystallized chert containing oxide-facies iron formation (Ano) was recognized in the felsic to intermediate volcanic suite. The occurrence is several tens of metres thick and has been traced along strike for at least 100 m. It consists almost entirely of quartz with minor graphite and centimetre-scale layers of near-monomineralic magnetite. The unit is intruded by both intermediate to mafic and felsic plagioclase-phyric dykes. Coarse garnet porphyroblasts form a garnet-rich reaction rim along the contact between the chert and felsic volcanic rocks.

3) A recrystallized clinopyroxene-bearing quartz-rich rock (Anc) in the Dougherty Lake and Lowe Bay areas (see accompanying map) occurs with calc-silicate rocks as thin ridges within rocks of the felsic to intermediate volcanic suite. Both occurrences can be traced along strike for up to two kilometres and appear several tens of metres thick. In addition to the coarse porphyroblastic clinopyroxene and more than 90 percent quartz, typical rocks contain minor magnetite and garnet. The unit is thought to represent a lean cherty iron formation of mixed oxide, carbonate and silicate facies.

Epiclastic sedimentary rocks generally occur as thin recessive sequences intercalated with rocks of the volcanic suite. Two units have been distinguished: 1) a heterogeneous layered sequence of mixed argillaceous to arenaceous biotite±hornblende±garnet±graphite gneisses (As) and, 2) a more homogeneous sequence of aluminous biotite-garnet-graphite±sillimanite wackes (Aw). The two are commonly interlayered on a scale of metres to hundreds of metres making distinction difficult. Both weather grey to rust and contain 10 to 30 percent leucosome.

Most samples from both units have a granoblastic texture with a grain size of about 1 mm; however, the grain size locally reaches up to 5 mm in rocks characterized by a granitic texture and lesser amounts of sillimanite and graphite. The fine- and coarse-grained varieties are interlayered on a scale of tens of centimetres to metres in some outcrops. Mutual contacts are generally gradational.

The mixed sedimentary rocks (As) differ in part from the more homogeneous biotite-rich aluminous wackes (Aw) by containing up to 35 percent hornblende. Minor cummingtonite is also present, and graphite forms up to 5 percent of some rocks. In the coarse-grained varieties, graphite may be the only means by which the rock can be related to a sedimentary precursor. The heterogeneity of these rocks suggests that they were derived from a variety of precursors, including calcic wackes, arenites and tuffaceous rocks.

The more homogeneous epiclastic biotite-garnet-graphite±sillimanite wackes (Aw) occur as layers tens to hundreds of metres in apparent thickness intercalated with rocks of the volcanic suites. They are about one millimetre in grain size and contain variable amounts of white biotite-garnet leucosome. Typical samples contain 20 to 30 percent biotite, 5 to 15 percent pink euhedral garnet and minor graphite. Sillimanite and tourmaline are minor constituents at many occurrences.

The aluminous wackes are similar to rocks making up much of the southern flank of the Kisseynew Domain (Ashton, 1987; Ashton *et al.*, 1987). Low-grade compositional equivalents also occur in the Welsh, Embury and File Lake areas of the Flin Flon Domain (Byers and Dahlstrom, 1954; Ashton *et al.*, 1987; Gilbert, 1988). They are thought to have been derived by sedimentary re-working of the volcanic component of the Amisk Group (Bailes, 1980).

c) Intrusive Rocks

Intrusive rocks, now for the most part orthogneisses and orthofoliates (Ashton and Leclair, 1990), constitute close to 50 percent of the Wildnest-Attitti Lakes area.

Granodioritic to Monzodioritic Rocks

Large bodies of gneissic granodiorite-tonalite (Gdh) occur throughout the area. Minor amounts of monzodioritic rocks (Dg) are spatially and perhaps genetically related to the gneissic granodiorite-tonalite in the Ewen and Attitti Lake areas (see accompanying map) .

The monzodioritic rocks (Figure 3) are medium grained, weakly foliated and occur as thin variably continuous units. Typical samples contain 20 to 50 percent hornblende and variable amounts of biotite, clinopyroxene, plagioclase, microcline and quartz.

The composition, appearance and spatial relationship of these rocks to the gneissic granodiorite-tonalites is similar to a suite of quartz monzonitic to dioritic intrusive rocks making up the Snake Rapids and Neagle Lake Plutons (Ashton, 1990b; Ashton *et al.*, in press) near the Flin Flon/Kisseynew Domain boundary.

The gneissic granodiorite-tonalite bodies (Gdh) occur as thin and laterally continuous units. The bodies are remarkably similar to each other, but display a large degree of internal heterogeneity. They exhibit a strong foliation which is generally parallel to that in the host supracrustal rocks. Typical outcrops consist of three rock types: the main intrusive rock, a 5 to 15 percent partial melt phase and mafic inclusions. The main intrusive rock is a medium-grained, grey granodiorite-tonalite (Figure 3) containing 5 to 15 percent hornblende and 5 to 15 percent biotite.

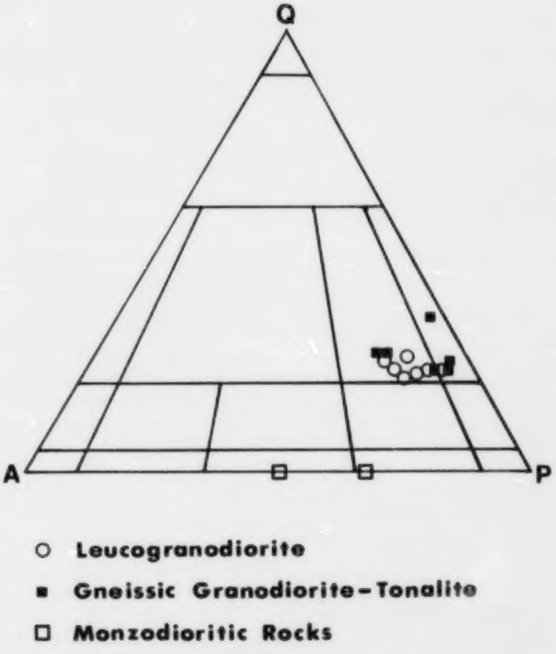


Figure 3 - Compositions of intrusive rocks based on the IUGS classification (compositional fields as outlined by Streckeisen, 1976).

The partial melt phase is coarser-grained, white and contains 5 to 10 percent hornblende and 0 to 5 percent biotite. It occurs as centimetre-scale concordant to crosscutting veins and pods which form a connected network. Blocks of the main intrusive rock are commonly rotated along the planes of the network. Other than in rare occurrences in the leucogranodiorite, the melt phase is restricted to the gneissic granodiorite-tonalite suggesting *in situ* derivation.

Mafic inclusions comprise up to 50 percent of some outcrops. Pyke (1961) used the variation in mafic inclusion abundance to distinguish a migmatite complex from his granodiorite-quartz diorite unit. This distinction has not been maintained here due to the irregular distribution of inclusions and the somewhat arbitrary unit constraints. In general, the mafic inclusions are most abundant towards the pluton margins. At least some of the inclusions are xenoliths of the Amisk intermediate to mafic volcanic suite but the coarser-grained and more quartzofeldspathic nature of others suggests that they represent early comagmatic phases of the granodiorite-tonalite plutons.

The gneissic granodiorite-tonalite rocks are interpreted as the high-grade metamorphic equivalents of the 1845 to 1860 Ma (Ansdell and Kyser, 1991) granodioritic suite in the Flin Flon Domain.

Leucogranodiorite

An extensive unit of leucocratic granodioritic rocks (Gdb) was distinguished from the gneissic granodiorite-tonalite on the basis of a lower colour index, higher biotite:hornblende ratio (hornblende may be absent), weaker foliation and by the tendency of mafic inclusions to occur as thin laterally continuous sheets comprising less than 10 percent of the outcrop (Figure 3). Leucogranodiorite bodies also tend to be more irregular in outcrop pattern. The weak internal foliation may be parallel or oblique to that of the host supracrustal rocks. Where it is oblique, it most commonly dips to the northeast, co-planar with the axial plane of the main map-scale folds (see accompanying map). Rare irregularly shaped mafic inclusions locally display a foliation at high angles to that in the leucogranodiorite.

Typical samples of leucogranodiorite are white, medium-grained, homogeneous and contain 5 to 15 percent biotite. Magnetite is a common accessory and muscovite is present rarely. In the western part of the area, hornblende is locally widespread and the rocks tend to weather pinker and contain abundant dykes and irregular segregations of a medium-grained to pegmatitic granodiorite phase.

The leucogranodiorite intrudes the supracrustal rocks and the gneissic granodiorite-tonalite. The relatively irregular outcrop pattern, weaker and locally younger foliation, and scarcity in the lower grade rocks to the east and south suggest that the leucogranodiorite is syntectonic.

Mafic-Ultramafic Intrusive Rocks

Mafic to ultramafic rocks (Um) occur as several small bodies in leucogranodiorite on the eastern side of the area and as one larger lens extending for more than a kilometre along the eastern side of Ewen Lake. A high aeromagnetic anomaly is associated with the Ewen Lake body (Geological Survey of Canada, 1970).

The mafic components are medium-grained rocks containing approximately equal portions of hornblende and plagioclase.

The ultramafic rocks exhibit knobby weathering due to medium- to coarse-grained porphyroblasts and are composed of variable proportions of olivine, orthopyroxene, clinopyroxene and amphibole with minor biotite and plagioclase.

The brown-weathering cores of the ultramafic bodies locally exhibit relict igneous cumulate textures. Primary orthopyroxene is locally preserved in these cores. Rare olivine may also be of igneous origin. The margins of the bodies are dark green to black clinopyroxene-bearing amphibolites.

The preservation of original igneous textures and the restricted occurrence in leucogranodioritic rocks suggest that the mafic to ultramafic rocks were emplaced during the later stages of deformation and metamorphism. Contacts tend to be complex but the mafic to ultramafic rocks appear to intrude the leucogranodiorite.

Deformed Medium-grained to Pegmatitic Granitoid Rocks

Deformed pink and white granitoid dykes and irregular segregations (Gp) occur throughout most of the area but rarely comprise more than about 15 percent of an outcrop. Most are concordant or near-concordant, medium-grained to pegmatitic, biotite-quartz-feldspar dykes of less than 1 m thickness. They were likely intruded over a range of time but many are thought to have been derived during and immediately following the peak of metamorphism, both as *in situ* and injected partial melts.

In a north-striking concordant zone 2 to 3 km west of Attitti Lake (Figure 2), gneissic granodiorite-tonalite (Gdh), leucogranodiorite (Gdb) and supracrustal host rocks contain 30 to 90 percent injected, pink, medium-grained to pegmatitic granitoid rocks (Gp). The eastern boundary of the zone is sharp and generally marked by a topographic high associated with the large volume of competent granitoid rocks. The western boundary is tentatively placed where the abundance of medium-grained to pegmatitic granitoid material drops to 10 to 30 percent, allowing easier recognition of the host rocks.

Well-developed to irregular layering ranges from a centimetre to several metres in thickness in the zone, and mafic inclusions and schlieren make up about 10 percent of most outcrops. The injected medium-grained to pegmatitic granitoid rocks contain 5 to 15 percent biotite and up to 5 percent magnetite. Hornblende and garnet occur in patchy distribution. Rare clinopyroxene porphyroblasts display rims of hornblende.

The rocks within this zone are variably deformed. In places, pegmatitic rocks have been disaggregated, and derived feldspar porphyroclasts display evidence of milling due to shearing. A weak northeast-plunging stretching lineation is locally developed and some quartz in the pegmatite is ribbony, displaying aspect ratios of about 5:1.

The abundance and sharp eastern contact of the medium-grained to pegmatitic granitoid rocks may result from injection along a developing structural discontinuity. Shearing probably accompanied emplacement and accounts for the local pegmatite disaggregation and ribbon quartz. A similar sheared pegmatite 6 km south of the study area yielded a U/Pb zircon age of 1806 ±2 Ma (Ashton *et al.*, in press).

Undeformed Medium-grained to Pegmatitic Granitoid Rocks

A less common set of weakly deformed to undeformed, pink, medium-grained to pegmatitic granitoid rocks was also recognized. They occur as concordant and crosscutting biotite-quartz-feldspar dykes and irregular segregations which are generally too small to appear on a 1:20 000 scale map. They may represent part of the ca. 1770 Ma Jan Lake Granite suite of granitic sheets and pegmatites (Macdonald and MacQuarrie, 1978; Bickford *et al.*, 1987; Chiarenzelli, 1989; Ashton *et al.*, in press).

d) Paleozoic Rocks

Two flat-lying blocks of unmetamorphosed and undeformed buff dolostone, each about 2 m thick by 5 m by 3 m, were recognized west of File Bay (UTM coordinates 667380E, 6106750N) and between Pearson and Rushton Lakes (UTM coordinates 664670E, 6100360N; see accompanying map). The bases of both blocks were unexposed, but because indicators of glacial ice flow imply south-southwest directed transport, the blocks are interpreted as small outliers of Paleozoic strata. Kupsch (1952) correlated similar rocks at the exposed unconformity 50 km to the south with the Ordovician Red River Dolomite.

3. Metamorphism

No metamorphic field gradient is obvious in the map area based on stable mineral assemblages. The leucosome of the gneissic granodiorite-tonalite is believed to be an *in situ* partial melt indicating that metamorphic temperatures exceeded the minimum melting curve. Sillimanite is the stable Al_2SiO_5 polymorph and apparent sillimanite-K-feldspar assemblages in the quartzofeldspathic gneisses suggest that conditions also exceeded the second sillimanite isograd. The absence of metamorphic orthopyroxene and garnet-cordierite assemblages constrain peak temperatures to upper amphibolite facies, likely in the 650° to 750°C range.

Peak pressures are more difficult to assess but the presence of sillimanite as the only Al_2SiO_5 polymorph sets limits of about 2 to 8 kb. The absence of cordierite in pelitic rocks indicates that peak pressures were towards the high end of that range. This is consistent with findings based on mineral stability observations and thermobarometric data from the area to the east and south (Digel *et al.*, this volume). That study shows a westward and northward increase in metamorphic grade towards the Wildnest-Attitti Lakes area from the east and south, and that peak conditions in the adjacent rocks were approximately 7 kb and 700°C.

Based on a separate reconnaissance thermobarometric study of the Reindeer Zone, Perkins (in press) suggests that peak pressures in the Attitti Block reached 8 to 11 kb. The absence of kyanite in the rocks mapped this summer seems inconsistent with such high values.

The wide variations in metamorphic grain size recognized in the sedimentary rocks are also common in the adjacent rocks east and south of the map area. The finest-grained rocks have a grain size of about one millimetre and an associated 10 to 30 percent concordant leucosome. The local development of a melanosome suggests that at least some of the leucosome was derived by *in situ* partial melting. In the coarser-grained rocks, grain sizes of 3 to 5 mm are common and the supracrustal rocks have a granitic texture. The composition, however, is not appreciably different from the finer-grained rocks. A separate leucosome phase is not obvious and minerals common in the finer-grained equivalents, such as sillimanite and graphite, appear less abundant. The grain size variations take place over a scale of tens of centimetres and it is common to find centimetre-scale layers of fine-grained material in the coarser equivalent. At present, the coarser-grained rocks are interpreted as diatexites resulting from near-complete anatexis of the finer-grained equivalents.

4. Structure

The Wildnest-Attitti Lakes area consists of a complexly folded central zone bounded by two north-striking straight zones. The eastern straight zone represents the attenuated western limb of the Wildnest Lake Synform[4] (Ashton *et al.*, 1987). The western zone is likely related to shearing associated with the Sturgeon-weir Shear Zone (MacQuarrie, 1979; Ashton and Wilcox, 1988). The folded central zone has undergone several phases of folding, some of which may be genetically linked to this regional shear deformation.

Primary features in the Wildnest-Attitti Lakes area are difficult to recognize as they have been largely obliterated by combined deformation and upper amphibolite facies regional metamorphism. They are limited to rare phenocrysts of quartz and feldspar in felsic to intermediate rocks interpreted as sub-volcanic porphyries.

a) D_1 Structures

The earliest and most prominent planar structural fabric is a generally well-developed pervasive foliation defined by compositional layering, lithological contacts and the preferred alignment of biotite and/or hornblende. Fold closures associated with this first phase of deformation (D_1) were not recognized.

b) D_2 Structures

Tight to isoclinal minor folds, intersection lineations and mineral lineations were developed during D_2 deformation. They tend to plunge moderately to the north-northeast in the north and gently to moderately to the southeast in the south due to later folding. Minor folds have northeast- to southeast-dipping axial planes. Two megascopic folds were developed during D_2 deformation. A doubly-folded sequence of supracrustal rocks south of Ewen Lake closes in the central Bentz Bay area (Figure 2). The closure is defined by mafic volcanic rocks which contain a thin traceable sulphide-facies iron formation. Numerous minor folds in the hinge zone plunge gently to the northeast.

The isoclinal Rushton Lake fold[5] (Figure 2) is also thought to be a D_2 structure based on the map pattern to the south (Ashton *et al.*, 1987). It plunges moderately to the southeast and has a moderately northeast-dipping axial plane.

c) D_3 Structures

The Bentz Bay Antiform and Ewen Lake Synform[6] are D_3 structures which both plunge gently to the northeast and have moderately northeast- to southeast-dipping axial planes (Figure 2; see accompanying map). Coeval minor folds and a strong linear fabric defined by hornblende, mineral clusters, rodding and rare mullion structures are particularly abundant in the hinge zones of the megascopic folds. Elsewhere, they are difficult to distinguish from D_2 structures due to their locally coaxial nature.

Two layer-parallel shear zones are variably developed along the eastern shore of Dezort Lake and west of Attitti Lake. Both are characterized by ribbon quartz and porphyroclastic K-feldspar produced from the dismembering of pegmatite dykes. The Dezort Lake shear zone consists of up to 200 m of protomylonitic gneissic granodiorite-tonalite. The shear zone west of Attitti Lake is marked by the extensive injection of deformed

---

(4) Wormworth Lake basin of Byers and Dahlstrom (1954)
(5) Pyke's (1961) Rushton Lake syncline
(6) Pyke's (1961) Ewen Lake syncline

medium-grained to pegmatitic granitoid rocks (Gp) into mainly gneissic granodiorite-tonalite. The shear zones are similar in appearance and orientation to structures within the main Sturgeon-weir Shear Zone to the west and southwest (Ashton *et al.*, 1987; Ashton and Wilcox, 1988). They may represent smaller related zones developed under the same stress regime. Rare northeast-plunging stretching lineations are consistent with the southwest-verging transport indicated in the Sturgeon-weir Shear Zone.

Folds which deform the sheared fabric in the Sturgeon-weir Shear Zone (Ashton and Wilcox, 1988) are thought to be coeval with D_3 folding in the Wildnest-Attitti Lakes area. On this basis, the shearing at Dezort Lake and west of Attitti Lake is tentatively considered part of the D_3 deformation.

d) D_4 Structures

Broad gentle warping of the eastern limb of the Bentz Bay Antiform and the regional north-south structural trend on the eastern side of the area is attributed to D_4 deformation. The gently east-plunging antiform[7] is responsible for rotating D_2 and D_3 minor folds into northeast-plunging orientations in the north and southeast-plunging orientations in the south.

e) D_5 Structures

Late mainly north-striking brittle faults (D_5) are defined by topographic lineaments (e.g. the fault through Attitti Lake described by Pyke, 1961). Rare exposures consist of altered cataclastic rocks displaying a strong fracture cleavage. Unexposed faults are inferred where offsets are apparent or where outcrops adjacent to marked lineaments contain abnormally abundant closed fractures lined with hematite, quartz and/or epidote. In the eastern part of the area, most brittle faults are layer parallel. The sense of displacement is unknown although some dip slip is inferred from the disappearance or thinning of units across the fault. The brittle faults generally crosscut the regional structure in the west and have an apparent left-lateral displacement. A significant dip-slip component is indicated on the Attitti Lake Fault by the absence of the leucogranodiorite in the western block, but the continuation of a weakly mineralized gossan zone suggests that the displacement is not too large (see accompanying map). The late brittle faults are probably related to a widespread set which includes the Ross Lake Fault, Granite Lake Fault, and late faulting along the Sturgeon-weir Shear Zone (Ashton, 1987; Ashton *et al.*, 1987).

5. Economic Potential

a) Sulphide Occurrences

The economic potential of the Wildnest-Attitti Lakes area is significantly increased with its affirmation as a high-grade equivalent of the volcanic facies of the Flin Flon Domain. Patchy oxide- and sulphide-facies iron formations indicate that the area was hydrothermally active prior to metamorphism. Apart from the known base metal deposits at Flin Flon, several showings (Coombe Geoconsultants Ltd., 1991) and occurrences of altered rocks (Ashton, 1987, 1989a; Ashton *et al.*, 1987) have been reported from high-grade rocks between Flin Flon and the study area.

Disseminated pyrrhotite is present throughout the volcanic rocks in the Wildnest-Attitti Lakes area, but is most common in the felsic to intermediate volcanic suite. Concentrations of greater than 3 percent pyrrhotite have been recorded on the accompanying map. Trace quantities of chalcopyrite are associated with several of these pyrrhotite concentrations and also occur as disseminated grains elsewhere in the volcanic suite.

b) Syn-volcanic Alteration

Fe-Mg Metasomatism

Garnet-anthophyllite-biotite rocks host or are spatially associated with most of the sulphide occurrences in the map area. Mg-rich rocks of this kind occur at virtually all volcanogenic massive sulphide deposits in amphibolite facies rocks of the Reindeer Zone (Ashton, 1979; Coombe Geoconsultants Ltd., 1991). Eskola (1914) pointed out that cordierite-anthophyllite rocks could not be derived by the isochemical metamorphism of unaltered, primary volcanic or sedimentary rocks. He suggested that such rocks were the result of Fe-Mg metasomatism. Akella and Winkler (1966) experimentally related cordierite-anthophyllite assemblages to chlorite-rich low-temperature equivalents by the reaction:

Al-chlorite + quartz = cordierite + gedrite + H_2O.

Various workers pointed out the spatial association of cordierite-anthophyllite rocks with volcanogenic massive sulphide deposits (Froese, 1969; De Rosen-Spence, 1969; Sangster, 1972; Bachinski, 1976). In the modern context, cordierite-anthophyllite rocks are viewed as the metamorphosed middle to upper amphibolite facies equivalent of rocks which have undergone Fe-Mg metasomatism.

Garnet-anthophyllite assemblages are also thought to result from alteration prior to metamorphism. They differ from cordierite-anthophyllite assemblages only by having a higher FeO:MgO ratio (Froese, 1969), and are also thought to result from Fe-Mg metasomatism. Garnet-anthophyllite assemblages have been reported from

---

(7) Pyke's (1961) Scholt Lake anticline

most sites of amphibolite facies massive sulphide mineralization in the Reindeer Zone including the Sherridon (Froese and Goetz, 1981) and Snow Lake (Froese and Moore, 1980) ore bodies in Manitoba, the Coronation mine east of Amisk Lake (Froese, 1969), the FON occurrence northwest of Neagle Lake and the Schott's Lake showing north of Wildnest Lake (Coombe Geoconsultants Ltd., 1991).

The patchy distribution, spatial relationship to sulphide concentrations and composition of the anthophyllite-garnet-biotite rocks in the map area suggest that they are also the metamorphosed equivalents of hydrothermally altered rocks which have been subjected to Fe-Mg metasomatism.

Aluminous Alteration and Silicification

The quartzofeldspathic gneiss (Ar) represents another compositional extreme not normally encountered in an unaltered sequence of dominantly volcanic rocks. It is spatially associated and in sharp contact with rocks of the felsic to intermediate and intermediate to mafic volcanic suites. This association, together with its irregular outcrop pattern, may indicate that the quartzofeldspathic gneiss represents altered equivalents of the felsic volcanic package. Several garnet-anthophyllite occurrences in or adjacent to the quartzofeldspathic gneiss indicate the altered nature of the sequence (see accompanying map). The sillimanite- and quartz-rich nature of the quartzofeldspathic gneisses could indicate that they are the products of aluminous alteration and silicification.

Silicification is reported from the alteration zones of many massive sulphide occurrences in the Reindeer Zone (Galley *et al.*, 1990). Relative enrichment in aluminum is also very common, but like Mg-metasomatism, it takes different forms at different metamorphic grades. Quartz-sericite schists host greenschist facies massive sulphide mineralization at the Hanson Lake Mine and McIlvenna Bay deposits, and andalusite is reported in the footwall alteration (Coombe Geoconsultants Ltd., 1991). In lower amphibolite facies rocks of the Snow Lake area, staurolite is recognized as a widespread indicator of aluminous alteration, and kyanite-sillimanite schists occur in the alteration zone of the Anderson Mine (Bailes *et al.*, 1987; Bailes, pers. comm., 1990). Staurolite, kyanite and sillimanite are also present in the alteration zone of the FON occurrence near Neagle Lake (de Tombe, 1988; Coombe Geoconsultants Ltd., 1991).

With the transition to higher grade, staurolite breaks down in the presence of muscovite according to the reaction (Thompson and Norton, 1968):

staurolite + muscovite + quartz = Al_2SiO_5 + garnet + biotite.

Since sillimanite is the stable Al_2SiO_5 polymorph throughout most, if not all, of the upper amphibolite facies rocks of the Reindeer Zone, an altered rock containing staurolite, muscovite and quartz at lower amphibolite facies would have to contain the assemblage sillimanite-garnet-biotite at higher grade.

There are a number of different reactions to describe the breakdown of sericite (a fine-grained form of mica generally made up of muscovite) but most also involve the production of Al_2SiO_5. One of the most common is the "second sillimanite isograd" (Evans, 1965; Althaus *et al.*, 1970):

muscovite + quartz = Al_2SiO_5 + K-feldspar + H_2O

which serves to illustrate that the higher-grade equivalent of a sericite schist is a sillimanite schist.

Since sillimanite-garnet-biotite and sillimanite-K-feldspar assemblages are generally thought to reflect sedimentary precursors, distinction of aluminous alteration in upper amphibolite facies rocks is difficult. It is noteworthy, however, that sillimanite is common in the quartz-rich gneisses which hosted the Sherridon ore bodies and is considered part of the alteration at the upper amphibolite facies Schott's Lake occurrence (Coombe Geoconsultants Ltd., 1991). The mineral assemblage sillimanite-garnet-biotite is also present at both of these localities.

Not all aluminous rocks can be treated as potential massive sulphide alteration zones. However, where they are associated with garnet-anthophyllite rocks in a volcanic package hosting numerous disseminated sulphide occurrences, such rocks deserve close examination. The quartzofeldspathic gneiss of the Gifford Bay area falls into that category.

6. Acknowledgments

C. Langen, I. Russell, D. Schreiner and D. Morrison provided cheerful and able assistance during the course of the fieldwork.

We are also grateful to D.J. Thomas for logistical assistance.

7. References

Akella, J. and Winkler, H.G.F. (1966): Orthorhombic amphibole in some metamorphic reactions; Contr. Min. Pet., v12, p1-12.

Althaus, E., Karotke, E., Nitsch, K-H. and Winkler, H.G.F. (1970): An experimental re-examination of the upper stability limit of muscovite plus quartz; Neues Jahrb. Mineral. Monatsh., p325-336.

Ansdell, K.M. and Kyser, T.K. (1991): Plutonism, deformation, and metamorphism in the Proterozoic Flin Flon greenstone belt, Canada: Limits on timing provided by the single-zircon Pb-evaporation technique; Geol., v19, p518-521.

Ashton, K.E. (1979): The geology of the Milton Island map sheet (east half), Saskatchewan (64D-10E); unpub. M.Sc. thesis, Univ. Sask., 131p.

__________ (1987): Preliminary geological map of the Kisseynew gneisses between the Sturgeon-weir River and Kisseynew Lake, Saskatchewan (parts of 63K-13, 63L-15,16, 63M-1, 63M-4); Geol. Surv. Can., Open File 1410.

__________ (1989a): The southern flank of the Kisseynew gneiss belt in Saskatchewan and its boundaries with the Flin Flon volcanic belt and the Hanson Lake Block; *in* Investigations by the Geological Survey of Canada in Manitoba and Saskatchewan during the 1984-1989 Mineral Development Agreements, Geol Surv. Can., Open File 2133, p17-20.

__________ (1989b): Kisseynew Project: Kisseynew Lake-Florence Lake; *in* Report of Activities 1989, Man. Energy Mines, p65-71.

__________ (1990a): The relationship between the Flin Flon volcanic belt and Kisseynew gneiss belt, Trans-Hudson Orogen, Saskatchewan (abstract); Geol. Assoc. Can.-Miner. Assoc. Can., Joint Annual Meeting, Vancouver, Prog. Abstr., v15, pA5.

__________ (1990b): Geology of the Snake Rapids Area, Flin Flon Domain (Parts of NTS 63L-9 and -10); *in* Summary of Investigations 1990, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 90-4, p4-12.

__________ (in prep.): Precambrian geology of the Mari Lake area, Saskatchewan; Geol. Surv. Can.

Ashton, K.E., Hunt, P.A. and Froese, E. (in press): Age constraints on the evolution of the Flin Flon volcanic belt and Kisseynew gneiss belt, Trans-Hudson Orogen, Saskatchewan and Manitoba; *in* Radiogenic Age and Isotopic Studies: Report 5, Geol. Surv. Can., Paper 91-2.

Ashton, K.E. and Leclair, A.D. (1990): Foliate: a useful term to complement the textural classification of foliated metamorphic rocks; Can. J. Earth Sci., v27, p1095-1097.

Ashton, K.E. and Wheatley, K.J. (1986): Preliminary report on the Kisseynew gneisses in the Kisseynew-Wildnest lakes area, Saskatchewan; *in* Current Research, Pt. B, Geol. Surv. Can., Pap. 86-1B, p305-317.

Ashton, K.E., Wheatley, K.J., Moser, D., Paul, D. and Wilcox, K.H. (1986): The Kisseynew gneisses of Saskatchewan: Update; *in* Summary of Investigations 1986, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 86-4, p145-148.

Ashton, K.E. and Wilcox, K.H. (1988): The Sturgeon-Weir shear zone: a thrusted boundary between the Flin Flon volcanic belt-Kisseynew gneiss terrain and Hanson Lake block, east-central Saskatchewan; Geol. Assoc. Can.-Miner. Assoc. Can.-Can. Soc. Pet. Geol., Joint Annual Meeting, St. John's, Nfld., Prog. Abstr., v13, pA3.

Ashton, K.E., Wilcox, K.H., Wheatley, K.J., Paul, D. and de Tombe, J. (1987): The boundary zone between the Flin Flon domain, Kisseynew gneisses and Hanson Lake block, northern Saskatchewan; *in* Summary of Investigations 1987, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 87-4, p131-134.

Bachinski, D.J. (1976): Metamorphism of cupriferous iron sulphide deposits, Notre Dame Bay, Newfoundland; Econ. Geol., v71, p443-452.

Bailes, A.H. (1980): Geology of the File Lake area; Man. Miner. Resour. Div., Geol. Rep. 78-1, 134p.

Bailes, A.H., Syme, E.C., Galley, A., Price, D.P., Skirrow, R. and Ziehlke, D.J. (1987): Early Proterozoic volcanism, hydrothermal activity, and associated ore deposits at Flin Flon and Snow Lake, Manitoba; Geol. Assoc. Can.-Min. Assoc. Can., Joint Annual Meeting, Field Trip Guidebook.

Bickford, M.E., Van Schmus, W.R., Collerson, K.D. and Macdonald, R. (1987): U-Pb zircon geochronology project: new results and interpretations; *in* Summary of Investigations 1987, Saskatchewan Geological Survey; Sask. Energy Mines, Misc. Rep. 87-4, p76-79.

Bruce, E.L. (1918): Amisk-Athapapuskow Lake district; Geol. Surv. Can., Mem. 105, 91p.

Byers, A.R. and Dahlstrom, C.D.A. (1954): Geology and mineral deposits of the Amisk-Wildnest Lakes area, Saskatchewan; Sask. Dep. Miner. Resour., Rep. 14, 177p.

Chiarenzelli, J.R. (1989): The Nistowiak and Guncoat gneisses: implications for the tectonics of the Glennie and La Ronge Domains, northern Saskatchewan, Canada; Ph.D. dissert., Univ. Kansas, 229p.

Coombe Geoconsultants Ltd. (1991): Base metals in Saskatchewan; Sask. Energy Mines, Open File Rep. 91-1, 218p.

De Rosen-Spence, A. (1969): Genesis of cordierite-anthophyllite rocks from Cu-Zn layers of the Noranda-Rouyn region; Can. J. Earth Sci., v6, p1339-1345.

de Tombe, J. (1988): The metamorphic stability fields of the Welsh Lake area in east-central Saskatchewan; unpubl. B.Sc. thesis, Queen's Univ., Kingston, 46p.

Eskola, P. (1914): On the petrology of the Orijarvi region in southwestern Finland; Bull. Comm. Geol. Finlande, no40.

Evans, B.W. (1965): Application of a reaction-rate method to the breakdown equilibria of muscovite and muscovite + quartz; Am. J. Sci., v263, p647-667.

Froese, E. (1969): Metamorphic rocks from the Coronation Mine and surrounding area; *in* Byers, A.R. (ed.), Symposium on the geology of Coronation Mine, Saskatchewan, Geol. Surv. Can., Pap. 68-5, p55-77.

Froese, E. and Goetz, P.A. (1981): Geology of the Sherridon Group in the vicinity of Sherridon, Manitoba; Geol. Surv. Can., Pap. 80-21.

Froese, E. and Moore, J.M. (1980): Metamorphism in the Snow Lake area, Manitoba; Geol. Surv. Can., Pap. 78-27.

Galley, A.G., Bailes, A.H., Syme, E.C., Bleeker, W., Macek, J.J. and Gordon, T.M. (1990): Geology and ore deposits of the Early Proterozoic Flin Flon and Thompson belts, Manitoba; Fieldtrip Guidebook, The International Association of the Genesis of Ore Deposits: 8th Symposium, Ottawa, 1990, 136p.

Geological Survey of Canada (1970): Attitti Lake, Saskatchewan; Geol. Surv. Can., Geophys. Pap. 4586 (aeromagnetic series).

Gilbert, H.P. (1988): Geological investigations in the Tartan Lake-Embury Lake area; *in* Report of Field Activities 1989, Man. Energy Mines, Miner. Div., p35-42.

Kupsch, W.O. (1952): Ordovician and Silurian stratigraphy of east central Saskatchewan; Sask. Dep. Miner. Resour., Rep. 10, 62p.

Macdonald, R. (1981): Compilation bedrock geology: Pelican Narrows and Amisk Lake areas (NTS 63M, 63L, part of 63N and 63K): Preliminary geological map; *in* Summary of Investigations 1981, Saskatchewan Geological Survey, Sask. Miner. Resour., Misc. Rep. 81-4, p16-23.

Macdonald, R. and MacQuarrie, R.R. (1978): Geological re-investigation mapping, Jan Lake area (part of NTS Area 63M); *in* Summary of Investigations 1978, Saskatchewan Geological Survey, Sask. Miner. Resour, Misc. Rep. 78-10, p16-24.

MacQuarrie, R.R. (1979): Geological re-investigation mapping, Birch Portage south, NTS area (63L-15S); *in* Summary of Investigations 1979, Saskatchewan Geological Survey, Sask. Dep. Miner. Resour., Misc. Rep. 79-10, p29-38.

McInnes, W. (1914): The basins of the Nelson and Churchill Rivers; Geol. Surv. Can., Mem. 30, 146p.

Perkins, D. (in press): Metamorphism of the Kisseynew gneisses and related rocks of the Reindeer Zone, Trans-Hudson Orogen, northern Saskatchewan; Can. J. Earth Sci., v28.

Pyke, M.W. (1961): The geology of the Attitti Lake area (west half), Saskatchewan; Sask. Dep. Miner. Resour., Rep. 54, 33p.

__________ (1966): The geology of the Pelican Narrows and Birch Portage areas, Saskatchewan; Sask. Dep. Miner. Resour., Rep. 93, 68p.

Sangster, D.F. (1972): Precambrian volcanogenic massive sulphide deposits in Canada: a review; Geol. Surv. Can., Pap. 72-22.

Satterly, J. (1932): Pelican Narrows area, Saskatchewan; Geol. Surv. Can., Summ. Rep. 1931, Pt C, p26-36.

Streckeisen, A. (1976): To each plutonic rock its proper name; Earth-Sci. Rev., v12, p1-33.

Taylor, F.C. (1958): Pelican Narrows, Saskatchewan; Geol. Surv. Can., Map 1-1958, with marginal notes.

Thompson, J.B. and Norton, S.A. (1968): Paleozoic regional metamorphism in New England and adjacent areas; *in* E-An Zen et al. (eds.), Studies of Appalachian Geology, Interscience Publisher (John Wiley and Sons), New York.

Wright, J.F. (1933): Amisk Lake, Saskatchewan; Geol. Surv. Can., Summ. Rep. 1932, Pt. C, p73-110.

Metamorphic P-T Investigations in the Southern Part of Kisseynew and Western Part of Flin Flon Domains[1]

Scott Digel[2], K.E. Ashton and K.H. Wilcox[3]

Digel, S., Ashton, K.E. and Wilcox, K.H. (1991): Metamorphic P-T investigations in the southern part of Kisseynew and western part of Flin Flon Domains; *in* Summary of Investigations 1991, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 91-4.

The Trans-Hudson Orogen (Figure 1) has seen a large increase in well-integrated detailed mapping, geochemistry, and geochronology over the last 10 years. As a result, our understanding of the geology of the area has improved to the point where geothermobarometry can address specific geological questions as well as provide information on a regional scale. This on going project was undertaken in 1990 to provide reconnaissance-scale pressure-temperature (P-T) data in parts of the Flin Flon and Kisseynew Domains (Figure 1). Future work will involve the analysis of samples from the Attitti and Hanson Lake Blocks (Figure 1) along the Wildnest-Tabbernor Transect (Ashton and Leclair, this volume).

Samples for geothermobarometry were collected during summer fieldwork (Ashton and Wheatley, 1986; Ashton *et al.*, 1986, 1987; Ashton, 1987, 1990). P-T results have been integrated with those obtained from rocks of the Annabel Lake area (Figure 2) by Wilcox (1990). Mineral analyses and P-T calculations were performed at the University of Calgary.

1. Methods

Mineral analyses were performed on a semi-automated ARL-SEMQ electron microprobe. Cores and rims in all minerals were analysed to check for compositional zoning. In general, only garnets were found to be zoned. Compositional data are available upon request from Digel.

Pressures and temperatures for metapelites were calculated using garnet-plagioclase-aluminosilicate-quartz (GPAQ) and garnet-biotite (Grt-Bt) equilibria with the database of Berman (1988), the computer program Ge0-Calc (Brown *et al.*, 1989), and the activity models in Ge0-Calc: plagioclase (Fuhrman and Lindsley, 1988), garnet (Berman, 1990) and biotite (McMullin, unpublished). Mineral abbreviations are given in Table 1. P-T results were produced from intersections of GPAQ and Grt-Bt equilibria in P-T space (Table 2).

In order to test the applicability to metamorphosed calc-alkaline plutons, pressures for three plutons were estimated using the aluminum-in-hornblende geobarometer developed by Johnson and Rutherford (1989) and Hollister *et al.* (1987).

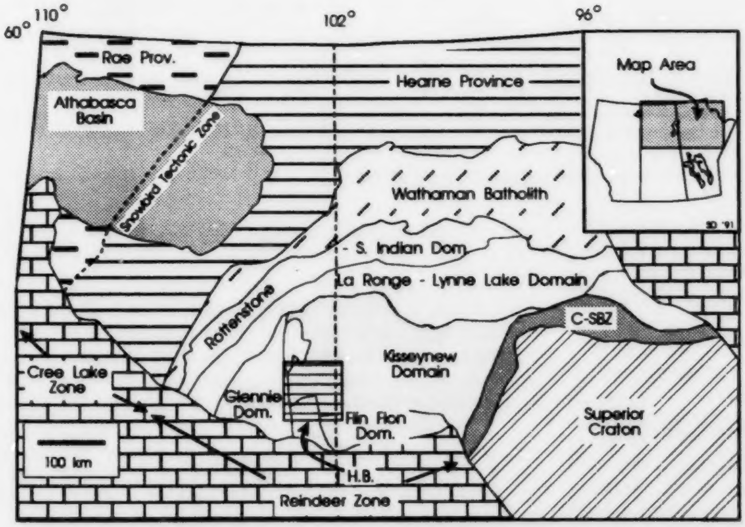


Figure 1 - Tectonic elements of the Trans-Hudson Orogen (modified from Bickford et al., 1990). Ruled box denotes area covered by Figures 2 to 4. Brick pattern represents extent of Paleozoic rocks. H.B. = Hanson Lake Block; C-SBZ = Churchill-Superior Boundary Zone.

2. Mineral Compositions

a) Metapelites

Biotite compositions are quite homogeneous in each sample. There is no significant compositional variation in the matrix between

---

[1] Saskatchewan Project A.121, from which this report derives, was funded in 1991 under the Canada-Saskatchewan Partnership Agreement on Mineral Development 1990-95.

[2] Department of Geology and Geophysics, University of Calgary, Calgary, Alberta, T2N 1N4

[3] Department of Earth Sciences, Carleton University and Ottawa-Carleton Geoscience Centre, Ottawa, Ontario, K1S 5B6

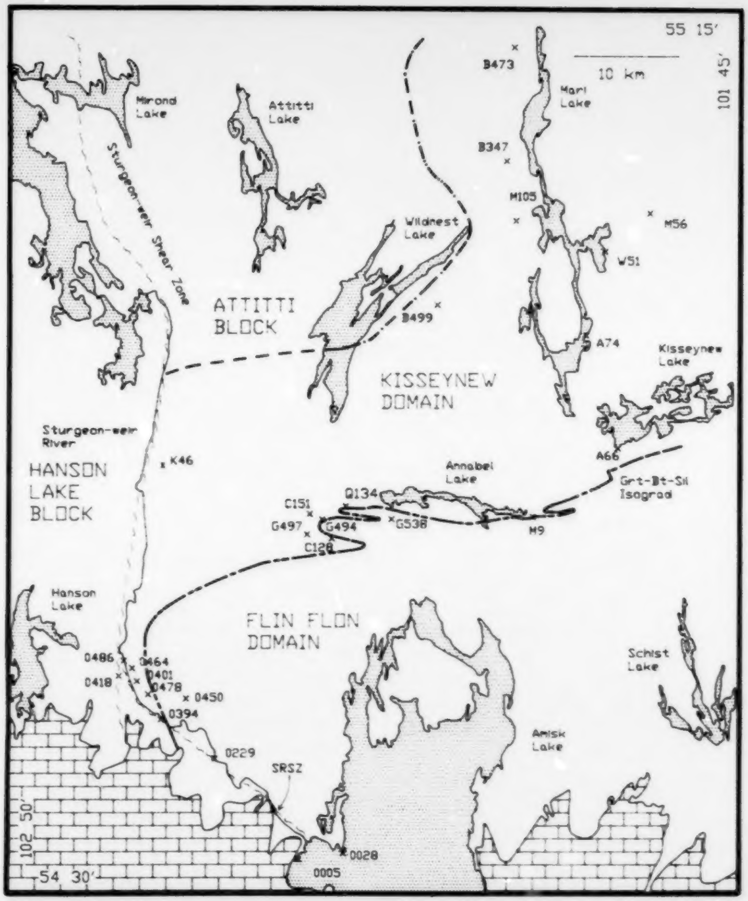

55 15′
10 km
101 45′
Mirond Lake
Attitti Lake
B473
Mari Lake
B347
M105
M56
Wildnest Lake
W51
Sturgeon-weir Shear Zone
ATTITTI BLOCK
B499
A74
Kisseynew Lake
KISSEYNEW DOMAIN
Sturgeon-weir River
HANSON LAKE BLOCK
K46
Annabel Lake
A66
Grt-Bt-Sil Isograd
Q134
C151
G497
G494
G538
H9
C128
Hanson Lake
FLIN FLON DOMAIN
Schist Lake
0486
0464
0418
0401
0478
0450
0394
0229
SRSZ
Amisk Lake
102 50′
54 30′
0028
0005

Figure 2 - Sketch map of the southern part of the Kisseynew and the western part of the Flin Flon Domains showing sample locations. SRSZ = Snake Rapids Shear Zone. Domain boundaries modified from Macdonald (1981).

grains adjacent to garnet and those from the matrix in general. Plagioclase compositions are fairly consistent within individual samples.

Unlike other metapelitic minerals, garnets are zoned. As garnet compositions are involved in both pressure and temperature estimates for metapelites, the evaluation of garnet zoning has a strong influence on the final P-T estimates. Compositional profiles were obtained for garnets in each sample. In general, two types of zoning were noticed. "Normal" growth zoning (Tracy, 1982), in which the spessartine (Mn-Al) component increases from rim to core, is typical of lower-grade rocks. "Reverse" zoning, in which the spessartine component decreases from rim to core, is typical of higher-grade rocks and results from retrograde modification of compositionally homogeneous garnets. The type of zoning was used to determine whether core or rim analyses should be applied to estimate P and T. For normally zoned garnets, rim compositions were used to estimate maximum P and T (Table 2). Core compositions were used for garnets exhibiting reverse zoning.

The grossular (Ca-Al) component is relatively constant in all garnet profiles, thus GPAQ equilibria are relatively unaffected by garnet zoning. As the Fe/Mg ratio differs in garnet cores relative to rims, the Grt-Bt Fe-Mg exchange equilibrium is the controlling factor in P-T estimates.

b) Calc-alkaline Plutons

Hornblende is unzoned in all samples except #0450 in which poikiloblastic cores are significantly more actinolitic than the rims. Zoning of hornblende is not unusual in calc-alkaline plutons (Hollister *et al.*, 1987). Rim analyses were used for sample #0450.

3. Pressure and Temperature Estimates

a) GPAQ

Although sillimanite is present in all of the hand samples, several of the polished thin sections are devoid of Al_2SiO_5 minerals. Those samples may not be saturated with respect to Al_2SiO_5 (i.e. activity less than 1). Accordingly, pressure estimates for these rocks represent maximum values (denoted by * in Table 2). Pressure and temperature estimates range from 2.7 to 7.3

Table 1. List of Mineral and Endmember Abbreviations (after Kretz, 1983).

Minerals

Bt	Biotite	Grt	Garnet
Ilm	Ilmenite	Kfs	K-feldspar
Ky	Kyanite	Mag	Magnetite
Pl	Plagioclase	Qtz	Quartz
Sil	Sillimanite	Spn	Sphene

Endmembers

alm	almandine	an	anorthite
grs	grossular	prp	pyrope
sps	spessartine		

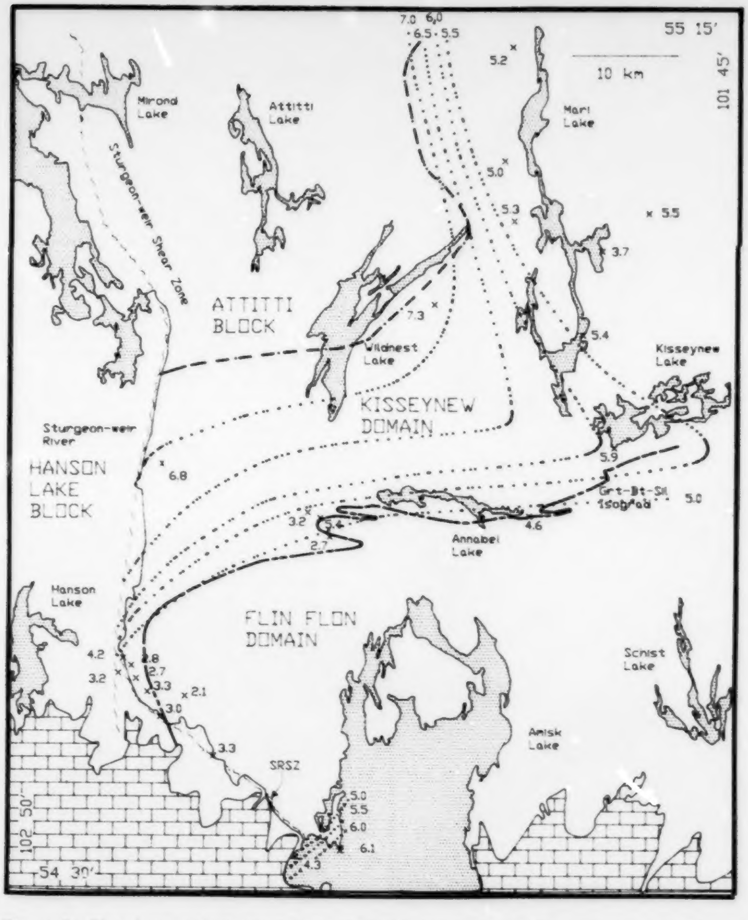

7.0
6.0
6.5
5.5
55 15'
5.2
10 km
101 45'
Mirond Lake
Attitti Lake
Mari Lake
Sturgeon-weir Shear Zone
5.0
5.3
5.5
3.7
ATTITTI BLOCK
7.3
Wildnest Lake
5.4
Kisseynew Lake
KISSEYNEW DOMAIN
Sturgeon-weir River
HANSON LAKE BLOCK
6.8
5.9
Grt-Bt-Sil Isograd
5.0
3.2
3.4
2.7
Annabel Lake
4.6
Hanson Lake
FLIN FLON DOMAIN
4.2
3.2
2.8
2.7
3.3
2.1
3.0
Schist Lake
Amisk Lake
3.3
SRSZ
5.0
5.5
6.0
6.1
4.3
102 50'
54 30'

Figure 3 - Sketch map showing contoured geobarometric results (symbols as in Figure 2).

kbar and 520°C to 725°C for metapelites (Table 2; Figures 3 and 4).

b) Aluminum in Hornblende

Hollister *et al.* (1987) describe the assemblage required to estimate pressure using Al^{IV} in hornblende. The critical assemblage from calc-alkaline rocks is Hbl-Bt-Spn-Ilm/Mag-An_{32}-Kfs-Qtz. Sample #0450 contains plagioclase with a composition of An_{14}. How this affects the estimated pressure is uncertain. Pressure estimates for calc-alkaline plutons between Amisk Lake and the Sturgeon-weir Shear Zone range from 2.1 to 4.3 kbar (Table 2, Figure 3).

4. Interpretation

a) Metapelites

Isograds based on metapelitic mineral assemblages illustrate a northward and westward increase in metamorphic gradient at the Flin Flon Domain-Kisseynew Domain and Flin Flon Domain-Hanson Lake Block boundaries, respectively (Ashton, 1990; in prep.). It has been suggested that the sillimanite-garnet-biotite isograd (Figure 3) be used to mark the boundary between the Flin Flon and Kisseynew Domains (Froese and Moore, 1980; *Zwanzig*, 1990). Several other isograds indicative of upper greenschist to middle amphibolite facies conditions have similar surface traces and plot to the immediate south. To the northwest, an increase in metamorphic grade can be locally inferred from textural gradations between fine-grained granoblastic material and their coarser, more granitic-textured equivalents and by the extensive development of leucosome. However, the absence of mineral phase changes makes it difficult to establish the extent and nature of these higher-grade areas and to quantify peak metamorphic conditions.

The P-T estimates from this study are generally consistent with the isograd pattern and with P-T constraints imposed from the stable mineral assemblages (Figures 3 and 4). All but 3 of the 20 samples for which pressure and temperature were determined yield values which are within the sillimanite-garnet-biotite stability field (Figure 5). The three samples which plot outside that field are assumed to have been altered and were ignored when contouring the results. All three come from the Welsh Lake area where they may have been affected by emplacement of the 1846 +14/-6 Ma Neagle Lake Pluton (Ashton *et al.*, 1987; in press).

Although the quantity of data is limited (Figures 3 and 4), several broad observations can be made:

1) There are at least two distinct components to the metamorphic gradient: a northward trend marking the Flin Flon Domain-Kisseynew Domain boundary, and a westward trend approaching the boundary between the Flin Flon and Kisseynew Domains to the east and the Hanson Lake and Attitti Blocks to the west.

2) Peak metamorphic conditions throughout much of the southern flank of the Kisseynew Domain were in the 5 to 5.5 kbar and 600 to 700°C range, but likely increased sharply to greater than 7 kbar and 700°C approaching the northern part of the Hanson Lake and Attitti Blocks. This is consistent with suggestions that the Flin Flon and Kisseynew Domains were

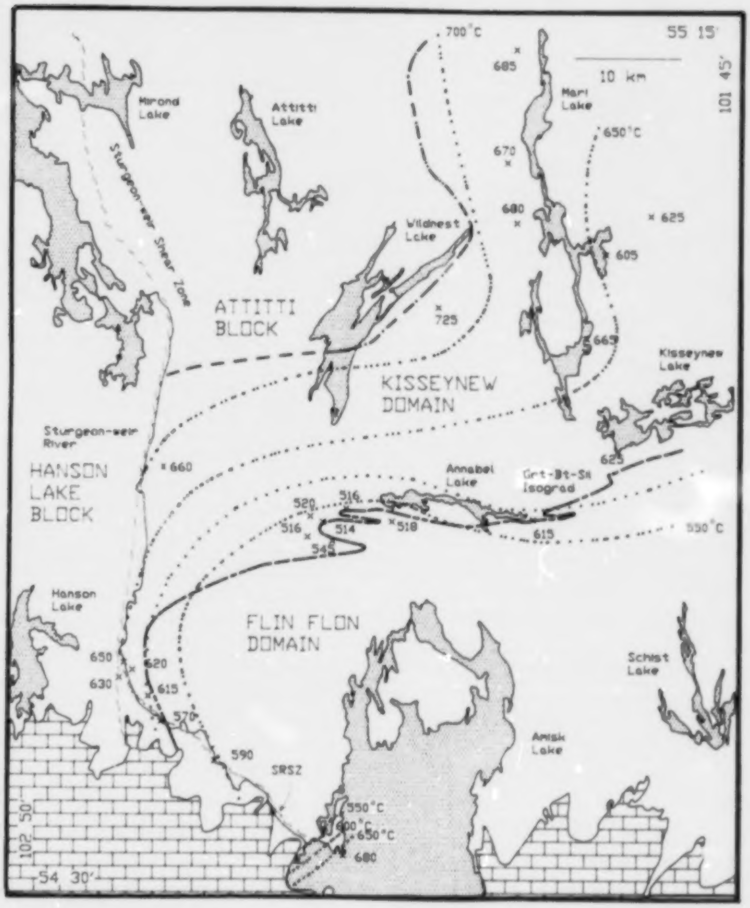

700°C
55 15'
10 km
101 45'
685
Mari Lake
650°C
670
680
625
605
Mirond Lake
Attitti Lake
Sturgeon-weir Shear Zone
Wildnest Lake
725
ATTITTI BLOCK
665
Kisseynew Lake
KISSEYNEW DOMAIN
Sturgeon-weir River
HANSON LAKE BLOCK
660
Annabel Lake
Grt-Bt-Sil Isograd
625
520
516
516
514
518
545
615
550°C
Hanson Lake
FLIN FLON DOMAIN
650
630
620
615
570
Schist Lake
Amisk Lake
590
SRSZ
550°C
600°C
650°C
680
102 50'
54 30'

Figure 4 - Sketch map showing contoured results of geothermometry (symbols as in Figure 2).

thrust southwestward over the Hanson Lake Block along the syn- to post-metamorphic Sturgeon-weir Shear Zone (Ashton and Wilcox, 1988). A discussion of the nature and role of the Attitti Block in this deformation may be found elsewhere in this volume (Ashton and Leclair, this volume).

3) East of Amisk Lake, the results are more complex. Temperature data are consistent with the isograds and define a thermal low flanked by steep gradients to the west and southeast (Figure 4; Ashton, 1990). The highest pressure results (Figure 3) come from samples yielding the highest temperatures suggesting a corresponding pressure low. The consistently low P-T values from samples of that area (Figure 5) could alternatively be related to their location along both sides of the Spruce Rapids Shear Zone (Ashton, 1990). Rare relict kyanite in sheared rocks from the zone suggests a relatively steep P-T path through the kyanite stability field. The low P-T values may not be consistent with such a prograde P-T path and could reflect retrograde effects associated with re-activation along the pre-existing Spruce Rapids Shear Zone.

b) Calc-alkaline Rocks

The results obtained from the aluminum-in-hornblende geobarometer are in close agreement with the metapelite data. This consistency may suggest that the low pressures obtained east of Amisk Lake are representative of peak metamorphic conditions. The aluminum-in-hornblende technique was developed, however, to determine the level of crustal emplacement of calc-alkaline plutons. More data are therefore deemed necessary to adequately evaluate the effects of subsequent metamorphism on pressure determinations.

5. Acknowledgments

T.W. Grover is thanked for help with the analytical work and numerous discussions.

6. References

Ashton, K.E. (1987): Preliminary geological map of the Kisseynew gneisses between the Sturgeon-weir River and Kisseynew Lake, Saskatchewan (parts of 63K-13, 63L-15,16, 63M-1, 63N-4); Geol. Surv. Can., Open File 1410.

__________ (1990): Geology of the Snake Rapids Area, Flin Flon Domain; *in* Summary of Investigations 1990, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 90-4, p4-12.

__________ (in prep): Precambrian geology of the Mari Lake area, Saskatchewan; Geol. Surv. Can.

Ashton, K.E. and Wheatley, K.J. (1986): Preliminary report on the Kisseynew gneisses in the Kisseynew-Wildnest lakes area, Saskatchewan; *in* Current Research, Pt. B, Geol. Surv. Can., Pap. 86-1B, p305-317.

Ashton, K.E., Wheatley, K.J., Moser, D., Paul, D., and Wilcox, K.H., (1986): The Kisseynew gneisses of Saskatchewan: update; *in* Summary of Investigations 1986, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 86-4, p145-148.

Ashton, K.E. and Wilcox, K.H. (1988): The Sturgeon-weir shear zone: a thrusted boundary between the Flin Flon volcanic belt-Kisseynew gneiss terrain and Hanson Lake block, east-central Saskatchewan; *in* Geol. Assoc. Can.-Min. Assoc. Can., Joint Annual Meeting, St. John's, Nfld., Prog. Abstr., v13, pA3.

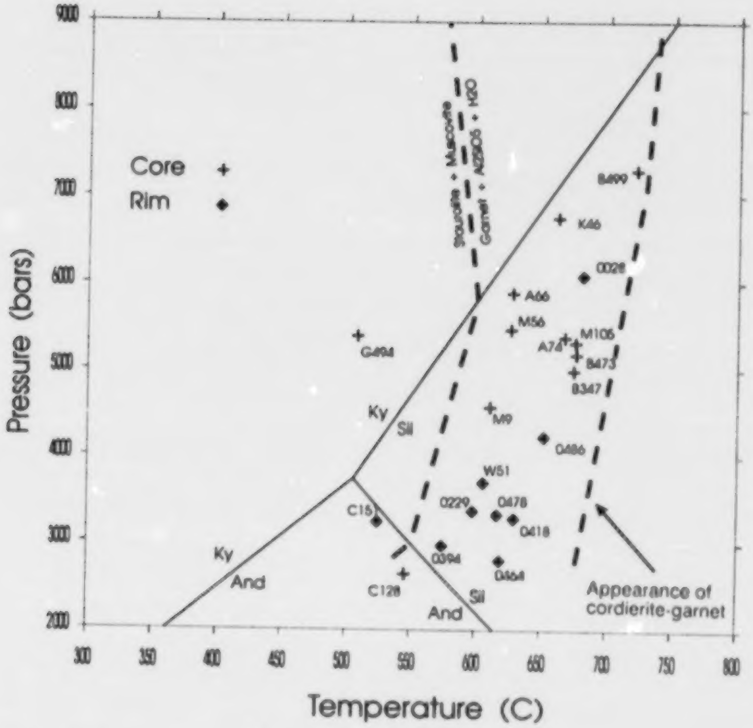


Figure 5 - Pressure-temperature diagram showing Grt-Bt and GPAQ intersections relative to the sillimanite-garnet-biotite stability field (Al_2SiO_5 reaction curves from Berman, 1988; others from Carmichael, unpublished).

Ashton, K.E., Wilcox, K.H., Wheatley, K.J., Paul, D., and de Tombe, J. (1987): The boundary zone between the Flin Flon and Kisseynew Domains, Kisseynew gneisses, and Hanson Lake Block, northern Saskatchewan; *in* Summary of Investigations 1987, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 87-4, p131-134.

Ashton, K.E., Hunt, P.A. and Froese, E. (in press): Age constraints on the evolution of the Flin Flon volcanic belt and Kisseynew gneiss belt, Trans-Hudson Orogen, Saskatchewan and Manitoba; *in* Radiogenic Age and Isotopic Studies: Report 5, Geol. Surv. Can., Pap. 91-2.

Berman, R.G. (1988): Internally-consistent thermodynamic data for stoichiometric minerals in the system Na_2O-K_2O-CaO-MgO-FeO-Fe_2O_3-Al_2O_3-SiO_2-TiO_2-H_2O-CO_2; J. Petrol., v29, p445-522.

__________ (1990): Mixing properties of the Ca-Fe-Mg-Mn garnets; Am.. Mineral., v75, p328-344.

Bickford, M.E., Collerson, K.D., Lewry, J.F., Van Schmus, W.R., and Chiarenzelli, J. (1990): Proterozoic collisional tectonism in the Trans-Hudson Orogen, Saskatchewan; Geol., v18, p14-18.

Table 2. Summary of Pressure (kbar)-temperature (°C) Estimates (bold print denotes values plotted and used for contouring on Figures 3 to 5).

* *Al_2SiO_5 in hand sample but not in microprobe mount*
n/a Garnets unsuitable for profiling due to fractured or poikiloblastic nature

GPAQ

	Sample	P_{Core}	T_{Core}	P_{Rim}	T_{Rim}
*	B473	5.2	685	3.3	600
*	M9	4.6	615	4.2	580
*	A74	5.4	665	3.1	575
	B347	5.0	670	2.4	530
*	M56	5.5	625	3.7	560
*	B499	7.3	725	6.1	665
	A66	5.9	625	6.0	585
*	C128	2.7	545	1.7	470
	K46	6.8	660	5.6	580
	M105	5.3	680	3.7	585
	C151	n/a	n/a	3.2	520
	W51	n/a	n/a	3.7	605
*	0486	3.9	625	4.2	650
*	0464	4.5	630	2.8	620
*	0028	6.3	695	6.1	680
	0478	n/a	n/a	3.3	615
	0418	3.5	650	3.2	630
	0229	n/a	n/a	3.3	590
	0394	n/a	n/a	3.0	570
	G494	n/a	n/a	5.4	514
	G497	n/a	560	n/a	516
	G533	n/a	538	n/a	518
	O134	n/a	575	n/a	516

Al in Hbl

Sample	P_1	P_2
0401	2.7	2.1
0450	2.1	1.7
0005	4.3	3.3

P_1: Hollister *et al.* (1987)
P_2: Johnson and Rutherford (1989)

Brown, T.H., Berman, R.G., and Perkins, E.H. (1989): PTA-SYSTEM: A Ge0-Calc software package for the calculation and display of activity-temperature-pressure phase diagrams; Am. Mineral., v74, p485-487.

Froese, E., and Moore, J.M. (1980): Metamorphism in the Snow Lake area, Manitoba; Geol. Surv. Can. Pap. 78-27.

Fuhrman, M.L., and Lindsley, D.H., (1988): Ternary feldspar modeling and thermometry; Am. Mineral., v73, p201-215.

Hollister, L.S., Grissom, G.C., Peters, E.K., Stowell, H.H., and Sisson, V.B. (1987): Confirmation of the correlation of Al in hornblende with pressure of solidification of calc-alkaline plutons; Am. Mineral., v72, p231-239.

Johnson, M.C. and Rutherford, M.J. (1989): Experimental calibration of the aluminum-in-hornblende geobarometer with application to Long Valley caldera (California) volcanic rocks; Geol., v17, p837-841.

Kretz, R. (1983): Symbols for rock-forming minerals; Am. Mineral., v68, p277-279.

Macdonald, R. (1981): Compilation bedrock geology: Pelican Narrows and Amisk Lake areas (NTS 63M, 63L, part of 63N and 63K): Preliminary geological map; *in* Summary of Investigations 1981, Saskatchewan Geological Survey, Sask. Miner. Resour., Misc. Rep. 81-4, p16-23.

Tracy, R.J. (1982): Compositional zoning and inclusions in metamorphic minerals; *in* Ferry, J.M. (ed.), Characterization of metamorphism through mineral equilibria, Rev. Mineral., v10, p355-397.

Wilcox, K.H. (1990): Geology of the Amisk-Welsh Lakes Area, Saskatchewan; unpubl. M.Sc. thesis, Univ. Calgary, 244p.

Zwanzig, H.V. (1990): Kisseynew gneiss belt in Manitoba: stratigraphy, structure, and tectonic evolution; *in* Lewry, J.J. and Stauffer, M.R. (eds.), The Early Proterozoic Trans-Hudson Orogen of North America, Geol. Assoc. Can., Spec. Pap. 37, p95-120.

Revision Bedrock Geological Mapping, Deschambault-Oskikebuk Lakes Area (Parts of NTS 73I-16, 63L-11, -12, -13 and -14)[1]

G.D. Delaney

Delaney, G.D. (1991): Revision bedrock geological mapping, Deschambault-Oskikebuk Lakes area (parts of NTS 73I-16, 63L-11, -12, -13 and -14); *in* Summary of Investigations 1991, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 91-4.

Revision mapping at a scale of 1:20 000 was completed this summer for a 390 km² southeast-trending strip between Jira Lake and Fisher Bay on Deschambault Lake (Figure 1). This work completes the selective remapping of supracrustal rocks in the Brownell, Wapawekka and Deschambault Lakes area in the southern part of the Glennie Domain (Delaney, 1988, 1989). Companion investigations have included geochronological studies (Delaney *et al.*, 1988, 1990) and a review of gold mineralization in the domain (Delaney, 1990a, b, 1991, in press). Objectives of the 1991 mapping were:

1) To better understand the character of the Oskikebuk Lake Group (Padgham, 1968), a supracrustal assemblage along the southern margin of the Oskikebuk Complex. Petrological, geochemical and geochronological investigations will be used to compare this succession with volcanic assemblages in the Brownell-Wapawekka Lakes area as well as elsewhere in the Glennie Domain and Trans-Hudson Orogen.

2) To document the character and extent of major high strain zones in the supracrustal rocks along the southern margin of the Oskikebuk Lake Complex. Mapping in 1988 and 1989 (Delaney, 1988, 1989) identified high strain zones in greenstones in the Brownell-Wapawekka Lakes area, including the economically significant major Hartley Shear Zone along the northern margin of the Oskikebuk Complex. This work also identified a splay of the Hartley Shear Zone along the southern side of the Oskikebuk Complex at the western, structurally tapered end of the complex. Brief reconnaissance examinations by Macdonald (1975) and Lewry (Lewry and Macdonald, 1988) at Oskikebuk Lake also identified highly strained rocks along the lake.

3) To investigate the economic potential of the area.

This work will lead to an enhanced understanding of the geological history of the southern Glennie Domain as well as aid in developing a model of the geodynamic setting of ore deposits which have formed there.

This report presents, in a preliminary form, the first specific descriptions of the geology of the area. Previously Padgham (1968) grouped together rock descriptions for supracrustal and granitoid rocks from throughout the Deschambault Lake District. The supracrustal rocks as well as some of the granitoids have been metamorphosed to amphibolite facies. For the purposes of simplicity, however, the prefix meta has been omitted in the following descriptions.

1. Previous Work

The first geological investigations were by McInnes (1913). DeLury (1926) completed a 1:126,720 scale geological map of the Wapawekka-Deschambault Lakes area. In 1963, Padgham (1966) mapped the northern-half of the Wapawekka Narrows (73I-16) map-area. Between 1963 and 1966 Padgham (1968) mapped the Deschambault Lake District, which included map sheets 63L-13, 63L-14 (W), and parts of 63L-11 and -12, at a scale of 1:63,360. Padgham's 1968 report contains a 1:31,680 map of the Maynard Lake area that includes the northern part of Oskikebuk Lake. The surficial geology of the area south and west of Wapawekka Lake was mapped by Langford (1973).

The westernmost part of the map area is included in the 1:250 000 compilation map of the bedrock geology of the Lac La Ronge area (Map 225A; Lewry and Slimmon, 1985). The eastern part is contained in a preliminary 1:250 000 compilation map for the Amisk Lake and Pelican Narrows areas (Macdonald, 1981).

An extensive reconnaissance lake sediment geochemical survey by the Geological Survey of Canada in 1974 (Hornbrook *et al.*, 1975, 1977, 1985) covered the present area.

Aeromagnetic coverage is provided by Geological Survey of Canada 1:63,360 scale aeromagnetic maps 4592G (73I-16), 4604G (63L-11), 4603G (63L-12), 4593G (63L-13) and 4594G (63L-14) and 1:250 000 scale aeromagnetic maps 7755G (73I) and 7754G (63L).

2. Regional Geology

The map area lies in the southernmost exposed part of the Glennie Domain, Trans-Hudson Orogen (Macdonald, 1987; Lewry and Collerson, 1990). The domain is a collage of narrow arcuate belts of supracrustal

---

(1) Saskatchewan Project A.122, from which this report derives, was funded in 1991 under the Canada-Saskatchewan Partnership Agreement on Mineral Development 1990-95.

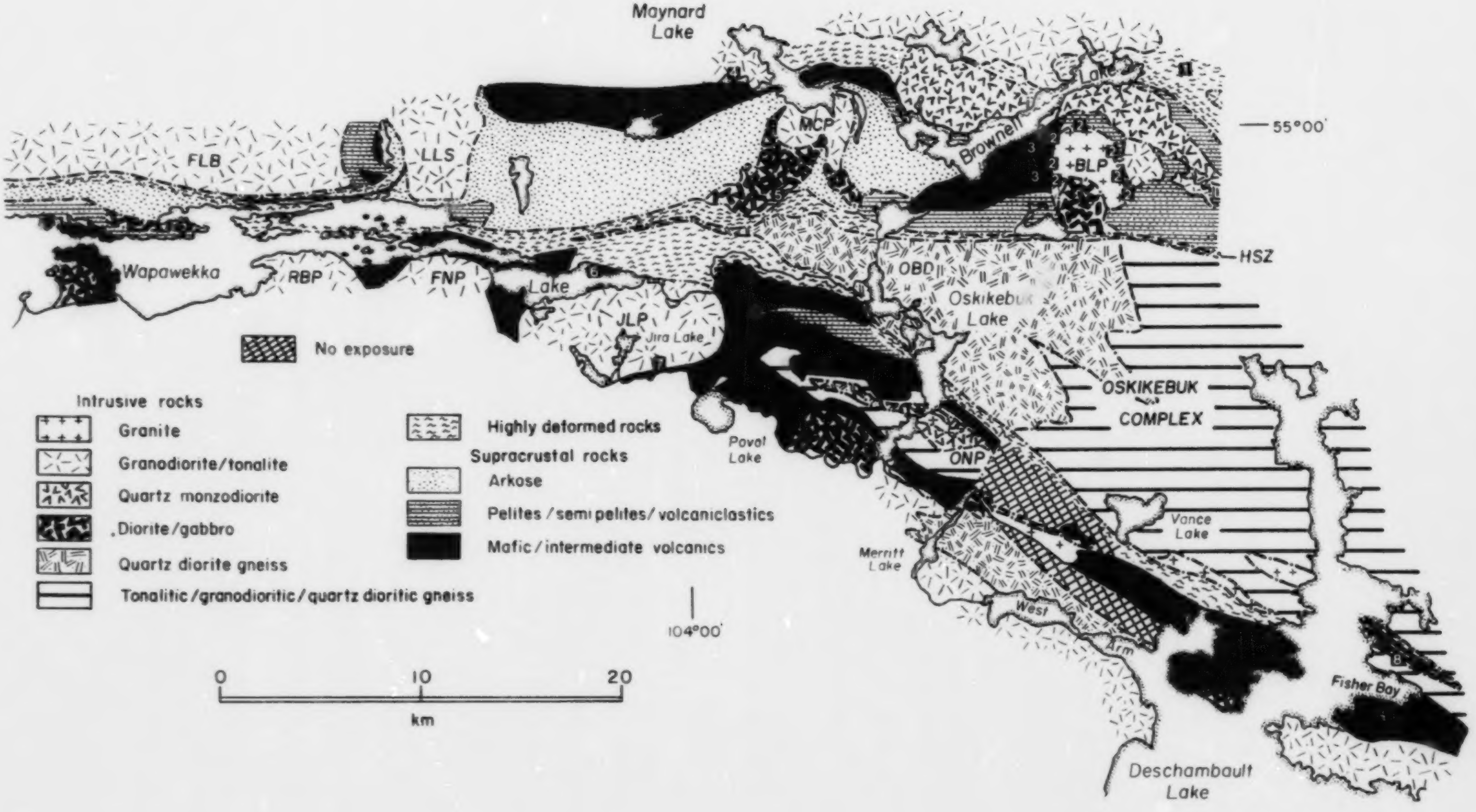

Maynard Lake
FLB
LLS
MCP
Brownell Lake
BLP
55°00'
HSZ
Wapawekka
RBP
FNP
Lake
JLP
Jira Lake
OBD
Oskikebuk Lake
OSKIKEBUK COMPLEX
Pavel Lake
ONP
Vance Lake
Merritt Lake
West Arm
Fisher Bay
Deschambault Lake
104°00'
No exposure
Intrusive rocks
Granite
Granodiorite/tonalite
Quartz monzodiorite
Diorite/gabbro
Quartz diorite gneiss
Tonalitic/granodioritic/quartz dioritic gneiss
Highly deformed rocks
Supracrustal rocks
Arkose
Pelites/semi pelites/volcaniclastics
Mafic/intermediate volcanics
0
10
20
km
1
2
3
4
5
6
7
8

Figure 1 - Geological sketch map of the Brownell, Wapawekka, Oskikebuk and Deschambault Lakes area. BLP, Brownell Lake Pluton; FLB, Folkerson Lake Batholith; FNP, Fisher Narrows Pluton; HSZ, Hartley Shear Zone; JLP, Jira Lake Pluton; LLS, Lacey Lake Stock; MCP, Maynard Creek Pluton; ONP, Oskikebuk Narrows Pluton, OBD, Oskikebuk Quartz Diorite Gneiss; RBP, Radar Bay Pluton. 1, Brownell Lake Property; 2, gold showings associated with the Brownell Lake Pluton; 3, gold showings in shear zones in Brownell Lake volcanics; 4 Alex showing; 5, M-10 Showing; 6, Wapa 13 showing; 7, Jira Lake showing, 8, Des 111 showing.

rocks in granitoids. Lewry (1981) suggested that the Glennie Domain was a remnant Archean microcontinent that collided with the La Ronge volcanic arc. Subsequent mapping and U-Pb dating (Chiarenzelli *et al.*, 1987; Chiarenzelli, 1989) has identified Archean age rocks in the Iskwatikan and Hunter Bay domes near the western margin of the Glennie Domain. As further detailed studies have been completed, the northern part of the domain has been interpreted as a stack of southeastward directed, refolded allochthonous thrust sheets formed during terminal collision in the Trans-Hudson Orogen (Lewry *et al.*, 1990). In order to further refine this model, a detailed geological transect of the southern Glennie Domain was commenced in 1988 (Delaney, 1988, 1989; Slimmon, 1988, 1989; Thomas, 1988). Lewry and Macdonald (1988, p. 40) in a synthesis of the early transect studies noted that "the southern part of the Glennie Domain appears to comprise a broadly homoclinal north-dipping stack of either predominantly supracrustal or plutonic sheets, typified by inhomogeneous strain and soled in some cases at least, by ductile high-strain zones. Upright cross-folding on predominantly northeast-trending axes modifies this pattern only superficially on the regional scale".

3. General Geology

A generally poorly-exposed assemblage of supracrustal rocks, the Oskikebuk Lake Group (Padgham, 1968), occurs in a southeast-trending belt from the east side of the Jira Lake Pluton to Fisher Bay on the west side of Oskikebuk Lake (Figure 1). The Oskikebuk Lake Group consists of variably altered, deformed and metamorphosed mafic to intermediate volcanics, subvolcanic intrusions, volcaniclastics and related immature sediments. Three major intermediate to basic intrusions occur in or adjacent to the supracrustal belt: 1) a serpentinized peridotite containing high nickel and chromium on the northeast side of the West Arm of Deschambault Lake, 2) a highly magnetic, xenolithic diorite between Merritt and Vance Lakes and 3) diorite on the southwest side of Oskikebuk Lake.

Supracrustals of the Oskikebuk Lake Group are bounded on the north by the Oskikebuk Complex - a mixture of granodioritic, tonalitic and quartz dioritic phases which are intruded by migmatitic granites. The Oskikebuk Hornblende Quartz Diorite Gneiss forms the tapered northwest end of the complex. A high strain zone of variable width and intensity, which has experienced multiphased deformations, is developed in supracrustal rocks along the southern side of the Oskikebuk Complex; high strain zones also occur within the complex in zones of lithologic heterogeneity. Along much of the south side the supracrustal belt is bounded by leucocratic tonalite of the Ballantyne Bay Pluton. North of the West Arm of Deschambault Lake, however, there is an older hornblende quartz diorite between the Ballantyne Bay Pluton and the supracrustals. Tonalitic and quartz dioritic gneisses form a major northwest-trending body in the west part of the Oskikebuk Lake Group. At Oskikebuk Lake, within this gneiss, is a small, late-stage zoned intrusion, the Oskikebuk Narrows Pluton. There is evidence of at least two major episodes of deformation.

Several electromagnetic conductors in the Oskikebuk Lake Group have been drill tested for massive sulphide mineralization. Although none of the drill holes intersected economically significant sulphides, core from one hole contained minor gold mineralization. Gold was recently discovered in quartz veins in shear zones along the southern margin of the Jira Lake Pluton.

4. Description of Lithologic Units

a) Oskikebuk Lake Group

The best exposures of the Oskikebuk Lake Group, and even these are limited, are in two areas: on the west side of Oskikebuk Lake and on the peninsula on the north side of the entrance to the West Arm of Deschambault Lake. Elsewhere exposure typically comprises a few scattered outcrops; in some areas such as on the south side of Vance Lake no outcrop was found. Because outcrop is so variable the following summary description of the Oskikebuk Lake Group is subdivided according to geographic locality.

West of Oskikebuk Lake

West of Oskikebuk Lake, on the north side of the tonalitic gneiss body in the Oskikebuk Lake Group (Figure 1), the group can be broadly subdivided into four main units that trend northwesterly. From north to south these are:

i) **Mafic to intermediate volcanic flows:** This subdivision consists of strongly foliated dark green amphibolite and weakly foliated to massive, medium-grained melanocratic to less commonly mesocratic hornblende granulite. Some of the amphibolitic units contain medium-grained phenocrysts or porphyroblasts of plagioclase. Porphyroblasts of hornblende are less common and amygdales are present only rarely. Small lenticular bodies of hornblendite occur in this sequence.

ii) **Wacke, volcaniclastic and minor intermediate flow rocks:** This unit encompasses a variety of rocks types that include the following: fine to medium-grained, leucocratic to mesocratic, strongly foliated, hornblende-plagioclase gneiss containing clots of fine- to medium-grained hornblende and biotite that compose 15 to 20 percent of the rock; fine-grained, strongly foliated, compositionally layered hornblende-plagioclase schist and gneiss which is locally calcareous; calcareous mesocratic hornblende-plagioclase schist; plagioclase porphyritic mesocratic hornblende-plagioclase schist and in less deformed sequences plagioclase phyric intermediate volcanic; wacke characterized by fine to medium-grained prismatic amphibole, composing 15 to 20 percent of the rock in a fine-grained matrix of quartz and plagioclase; leucocratic hornblende-quartz-plagioclase schist containing abundant pebble-size felsite clasts.

iii) **Mixed metasediments:** This unit comprises a variably altered sequence of quartzofeldspathic sediments and minor intermediate to felsic volcanics and volcaniclastics that include: biotitic quartzofeldspathic schist and gneiss; strong to very strongly foliated quartzofeldspathic schist and gneiss; conglomeratic schists characterized by variable amounts of pebble-sized, fine-grained quartzofeldspathic clasts; amphibole-biotite-quartzofeldspathic schists composed of 5 to 7 percent biotite with minor disseminated amphibole; compositionally layered hornblende-plagioclase gneiss; calc-silicate-bearing quartzofeldspathic schist with porphyroblasts of plagioclase flattened parallel to foliation; siliceous schist.

iv) **Mafic to intermediate volcanics:** This unit comprises: dark grey to dark greenish-grey amphibolite which to the northeast of Wapawekka Lake and along the Deschambault River is very strongly foliated with a well developed fissility; foliated and lineated, calcareous mesocratic hornblende-plagioclase schist; garnet porphyroblastic, leucocratic to mesocratic, hornblende-plagioclase schist; dark green, medium- to coarse-grained hornblendite; melanocratic uralite schist; hornblende porphyroblastic schist to medium-grained melanocratic hornblende diorite.

West of Oskikebuk Lake, south of the tonalitic gneiss is a narrow belt of strongly foliated, mesocratic to melanocratic, hornblende-plagioclase schist and gneiss as well as amphibolite. North of Povol Lake a buff to salmon colored felsite, composed of approximately 50 percent quartz, locally contains abundant fine-grained, grey xenoliths composed of various amounts of amphibole, biotite, quartz and feldspar. Contacts between the felsite and adjacent supracrustals are diffuse; locally there are blotchy patches of biotite that contain abundant coarse-grained porphyroblasts of garnet. North of Povol Lake, near the contact with the Jira Lake Pluton, exposure is poor and there are only a few scattered outcrops of highly fissile amphibolite.

Peninsula, West Arm, Deschambault Lake

Much of the peninsula on the northeast side of the entrance of the West Arm to Deschambault Lake is underlain by dark green and fine-grained mafic to intermediate volcanics. The main varieties are plagioclase phyric intermediate to mafic flows, and amphibolite and hornblende-plagioclase granulite which are less common. Amygdales and in rare cases pillows are preserved in flows. Locally, such as along Deschambault Lake, the flow rocks are intruded by subvolcanic diorite dykes and sills. Rare intermediate flow breccia and volcaniclastic fragmentals are present in the sequence. On the north side of the peninsula fine-to medium-grained leucocratic to mesocratic gneiss are more common; some of these are probably volcaniclastics. Also towards the north side of the peninsula there are lenticular masses of strongly foliated to mylonitized quartzofeldspathic gneiss, some of which are characterized by medium-grained quartz "eyes" in a fine-grained quartz-feldspar matrix.

Other Localities

East of Oskikebuk Lake the Oskikebuk Lake Group is segmented by a composite granitoid that includes quartz porphyroblastic tonalitic gneiss and the Oskikebuk Narrows Pluton. The group here is only exposed in a few scattered outcrops, most of which consist of variably feldspathized and granitized amphibolite. Near the contact with Oskikebuk Lake Complex there are amphibolitic and quartzofeldspathic mylonites.

On the south side of Fisher Bay, where inland exposure is poor, the Oskikebuk Lake Group includes: medium-grained mesocratic hornblende granulite, massive to compositionally layered dioritic gneiss, amygdaloidal mafic volcanic, and amphibolite. Within these rocks are a series of foliation-parallel layers of strong to very strongly foliated, salmon-coloured plagioclase- porphyroblastic granite to granodiorite.

Approximately one kilometre north of Fisher Bay is a highly tectonized and migmatized supracrustal sequence characterized by intervals a few metres to tens of metres thick of compositionally layered, fine-grained rock composed of various proportions of hornblende, quartz and plagioclase. Some of the hornblende-rich layers have been boudinaged. Between the compositionally layered intervals foliation-parallel layers and lenses of light grey, fine- to medium-grained leucocratic granodiorite with a very strong subhorizontal stretching lineation occur.

b) Granitoid Rocks

Diorite and Peridotite

Three significant intermediate to basic intrusions occur in the map area: 1) the diorite lying between Merritt Lake and Vance Lake which includes a highly magnetic xenolithic phase, 2) the diorite on the southwest side of Oskikebuk Lake and 3) the serpentinized peridotite on the northeast side of the entrance to the West Arm of Deschambault Lake.

Diorite - Merritt Lake

The northeast side of Merritt Lake is underlain by a bulbous-shaped, composite basic intrusion approximately 1 km in diameter. On the east side, the intrusion tapers abruptly into a narrow irregular tail that has been traced to the south side of Vance Lake. The main phase is a dark grey, leucocratic to mesocratic diorite to quartz diorite comprising up to 80 percent xenoliths. Xenoliths range in size from a couple of centimetres to 60 cm and consist of a variety of rock types: amphibolite; mesocratic to melanocratic diorite or gabbro; plagioclase phyric intermediate volcanic; fine- to medium-grained, mesocratic to melanocratic hornblende granulites; and fine-grained, compositionally layered, hornblende-plagioclase volcaniclastics. The matrix to the xenoliths contains clots of medium- to coarse-grained hornblende up to one metre wide. The unit as a whole has anomalously high magnetic susceptibility compared to all other rocks in the map-area and is defined by a distinct high magnetic anomaly on Federal-Provincial Aeromagnetic Map 7754G.

A more geographically restricted phase of the basic intrusion between Merritt Lake and the south side of Vance Lake comprises dark green diorite to gabbro containing 20 to 40 percent medium-grained porphyroblasts of hornblende. A minor phase, only observed along the shore of Merritt Lake, comprises a medium- to coarse-grained hornblendite.

Diorite - Southwest of Oskikebuk Lake

A large, generally poorly exposed, but apparently multiphased, basic intrusion underlies the southwest side of Oskikebuk Lake. Phases within this intrusion include: dark grey, massive, medium-grained diorite containing 30 to 40 percent hornblende; buff, massive to foliated, medium-grained diorite to quartz diorite containing approximately 15 percent hornblende and biotite; dark grey, fine-grained, mesocratic tonalite to quartz diorite containing scattered medium-grained porphyroblasts of hornblende; and dark grey, medium-grained, mesocratic hornblende tonalite.

Serpentinized Peridotite

A 1.5 km long by 800 m wide body of serpentinized peridotite occurs in mafic to intermediate volcanics of the Oskikebuk Lake Group on the peninsula on the northeast side of the entrance to the West Arm of Deschambault Lake.

Oskikebuk Complex

The Oskikebuk Complex comprises a mixture of granodioritic, tonalitic and quartz dioritic rocks and later migmatitic felsic granitoids along the north side of supracrustal rocks (Figure 1). Two subunits have been recognized: 1) the Oskikebuk (hornblende) Quartz Diorite Gneiss forming the tapered northwest end of the complex, and 2) the eastern part.

Oskikebuk Quartz Diorite Gneiss

The Oskikebuk Quartz Diorite Gneiss is grey to dark grey, medium grained and composed of 12 to 20 percent hornblende. Much of this unit contains coarse-grained plagioclase porphyroblasts and locally there are coarse-grained porphyroblasts of hornblende. The east-trending Hartley Shear Zone (Delaney, 1988, 1989) lies along the northern side of the quartz diorite gneiss. Along the northern margin of the unit augen-shaped plagioclase porphyroclasts are common. West-northwest of Oskikebuk Lake the unit has been structurally thinned to a 100 to 200 m thick "tail" in outcrop surface which can be traced another 6.5 km to the west. In this area a belt of highly deformed rocks mantle the Oskikebuk Quartz Diorite Gneiss on both the north and south sides.

In the vicinity of Oskikebuk Lake, the southern 3 km of the Oskikebuk Quartz Diorite Gneiss has been intruded by sills, dykes and veins of pink, fine-grained, leucocratic tonalite containing a few percent biotite. These range in thickness from a few centimetres to several metres. Some exhibit microscopic rodding which defines a stretching lineation. Within the leucocratic tonalite phase are medium-to coarse-grained quartz-feldspar veins and veinlets.

Eastern Part of Oskikebuk Complex

The eastern part of the Oskikebuk Complex consists of leucocratic to rare mesocratic tonalite, granodiorite and quartz diorite, all typically plagioclase porphyroblastic. A U-Pb zircon date of 1850 ±4 Ma (Bickford *et al.*, 1986) was obtained for this unit from a tonalite collected along the narrows in Deschambault Lake. The complex is intruded by late stage granitic dykes, sills and veins.

Granitic Migmatites

In addition to numerous late-stage granitic dykes, sills and veins, particularly in rocks of the Oskikebuk Complex, there are three large late-stage granitic bodies of a migmatitic origin. One of these, unit GM-3, located between Vance Lake and the West Arm of Deschambault Lake is a pinkish-buff, medium-grained granite to granodiorite containing between 12 and 20 percent biotite and hornblende. The rock varies from moderately to very strongly foliated. Xenoliths of mesocratic intermediate intrusive rocks are common and rare xenoliths of layered hornblende volcaniclastic also occur. Folded and boudinaged veins of pink quartz-feldspar pegmatite are present in this intrusion.

The other two large granitic migmatite bodies occur east of Vance Lake. The more southerly of these (unit GM-1) is a buff to salmon, fine- to medium-grained granite containing as much as 5 percent biotite, mostly as disseminated fine grains but also in minor medium-to coarse-grained porphyroblasts. The rock also contains minor euhedral, fine- to medium-grained garnet locally. The more northerly of the bodies is a pink, fine- to medium-grained, leucocratic granodiorite to granite containing a few percent disseminated biotite. This rock also occurs as dykes and sills in tonalitic to dioritic rocks that flank the granitic core.

Quartz Porphyroblastic Tonalitic Gneiss

A tonalitic gneiss complex occurs in the Oskikebuk Lake Group between the northeast end of Povol Lake and the north side of Merritt Lake. Much of this body is a fine- to medium-grained tonalite with millimetre-to centimetre-size, lineation-parallel clots of grey quartz and a few percent fine-grained biotite in foliation-parallel clots. Elsewhere, such as on the west side of Oskikebuk Lake and in the mantle of this unit on the north side of the Oskikebuk Narrows Pluton, the most common phase of the tonalitic gneiss contains pink, centimetre-size, oovoid-shaped porphyroblasts of quartz. On the west side of Oskikebuk Lake the unit contains a buff hornblende tonalite to quartz diorite phase that contains between 12 and 20 percent hornblende.

West of Oskikebuk Lake the tonalite gneiss contains lenses of medium-grained, dark green hornblendite. Four of these ultramafic lenses have been mapped; the largest, located about 2 km east of the southern narrows on Oskikebuk Lake, is a tear-shaped body approximately 1000 m long by 300 m wide.

Oskikebuk Narrows Pluton

The Oskikebuk Narrows Pluton (Figure 1; Padgham, 1968) is a small, elliptical-shaped, zoned intrusion. The core is a buff to salmon, leucocratic, microcline porphyritic granite and granodiorite which is quite similar to the microcline porphyritic granite phase of the Brownell Lake Pluton (Padgham, 1968; Delaney, 1988). The external part of the pluton is a dark grey, medium-grained mesocratic hornblende monzodiorite. Although Padgham (1968) noted the similarity of this pluton to the Maynard Creek and Brownell Lake Plutons, the zoned character of the Oskikebuk Narrows Pluton was not previously recognized. U-Pb age determinations on zircons from the felsic phases of the Brownell Lake and Maynard Creek Plutons have yielded dates of 1831 ±9 Ma and 1832 +9/-3 Ma respectively (Delaney *et al.*, 1990).

Ballantyne Bay Pluton

Along much of the south side, the Oskikebuk Lake Group is bounded by leucocratic tonalite of the Ballantyne Bay Pluton. South of Fisher Bay the main phase of the pluton is a buff to salmon, fine- to medium-grained tonalite containing 2 to 3 percent disseminated fine-grained biotite. This phase is cut by felsite and pegmatite veins. Between Deschambault and Oskikebuk Lakes, phases in the Ballantyne Bay Pluton include: pinkish-buff, medium-grained, leucocratic tonalite containing 3 to 5 percent biotite in foliation parallel clots; pinkish-buff, fine- to medium-grained tonalite containing no mafic minerals; and fine- to medium-grained, buff tonalite with 3 percent disseminated fine-grained biotite.

The north side of the West Arm of Deschambault Lake is underlain by grey to buff grey, medium-grained, massive to more commonly plagioclase porphyroblastic, leucocratic hornblende-quartz diorite. This unit, which is similar to some phases in the Oskikebuk Complex, is intruded by the Ballantyne Bay Pluton.

Jira Lake Pluton

The Jira Lake Pluton lies on the southern side of the east end of Wapawekka Lake. The northern margin of the intrusion was mapped in 1989 (Delaney, 1989) and the southeastern part during the current study. Three distinctive phases similar to those mapped by Padghan (1966) are distinguished:

1) The northern margin consists of a buff to buff grey, massive to foliated, leucocratic granodiorite to tonalite that weathers with a distinctive granular texture. Plagioclase ranges from 4 to 7 mm in size, is anhedral to subhedral and set in a finer-grained matrix of quartz, plagioclase and biotite that composes from 7 to 12 percent of this phase. Locally, quartz has a distinctive salmon colour.

2) Much of the rest of the eastern and southern parts of the pluton is a light grey, medium-grained, equigranular to porphyroblastic tonalite containing from 7 to 15 percent biotite and hornblende in various proportions; rare phases contain 17 to 20 percent hornblende.

3) The core of the Jira Lake Pluton is a coarser tonalite containing up to 7 percent biotite.

5. Structure

a) High Strain Zones

Previous mapping (Delaney, 1988, 1989) identified a major structural break, the east-trending Hartley Shear Zone, along the northern side of the Oskikebuk Quartz Diorite Gneiss. At the western end of the structurally tapered tail of the gneiss the Hartley Shear Zone merges with another apparently wide but poorly exposed zone of highly deformed rocks. During the current mapping the continuation of this high strain zone was traced along the southern margin of the Oskikebuk Complex, although a full appreciation of its character and extent was severely hampered by poor exposure. The highly strained rocks along the southern margin of the Oskikebuk Complex include the highly fissile mylonitic amphibolites west of Oskikebuk Lake and the very strongly foliated quartzofeldspathic gneisses and schists north of the West Arm of Deschambault Lake. These are interpreted as derived from a variety of protoliths. West of Oskikebuk Lake high strain zones occur throughout the supracrustal sequence, forming rocks such as very strongly foliated and highly fissile mylonitic amphibolites.

b) Structural History

Preliminary analysis of structural data reveals the presence of two major episodes of deformation as well as other minor events. The first event (D1) formed a pervasive, prominent, southeast-striking foliation that generally dips moderately to steeply north and has a strong subhorizontal mineral lineation. This lineation is coaxial with rare small-scale folds. The first deformation also formed the major ductile shear zone and associated folds along the southern margin of the Oskikebuk Complex, as well as heterogeneous shear zones throughout the supracrustal assemblage.

The second episode of deformation (D2) formed a crenulation cleavage which varies in strike from north-northwesterly to north-northeasterly and generally dips steeply to the east.

Structures attributed to these two events have been previously recognized in the Brownell and Wapawekka Lakes areas. The D1 event is correlated with the event in the Brownell Lake area which formed the overturned Brownell Lake Syncline and the Carroll Lake antiform (Delaney, 1988). Major east-trending high-strain zones such as the Hartley Shear Zone and the unnamed one along the margin of the Folkerson Batholith are also associated with D1.

The D2 event is correlated with the prominent D2 fracture cleavage and locally crenulations or small-scale northeast- to north-trending folds in the Wapawekka Lake area. Farther east in the Maynard and Brownell

Lakes area, deformation attributed to this event formed north- to northwesterly-trending folds including the Maynard Creek Anticline, the Brownell Lake Syncline and the Brownell Lake Anticline.

6. Economic Geology

A review of the Mineral Assessment Files of Saskatchewan Energy and Mines reveals that much of the Oskikebuk Lake Group between Jira Lake and Fisher Bay on Deschambault Lake has been investigated for massive sulphide potential. Airborne and ground electromagnetic surveys have identified a number of conductors of variable strength, commonly parallel to the trend of the belt. Some of these conductors were tested by diamond drilling but no economically significant sulphide intersections have been reported although minor gold mineralization was intersected in one hole (Des 111, Figure 1). More recently, shear zone-hosted gold mineralization was discovered at Jira Lake. The serpentinized peridotite on the north side of the West Arm to Deschambault Lake was reported to have a high nickel content (Padgham, 1968).

a) Gold

i) **Des 111**: Mineralization was intersected in a diamond drill hole, Des 111, collared on the north side of Fisher Bay, Deschambault Lake (UTM co-ordinates 599650 m E, 6068460 m N). Between March 1972 and January 1973 Hudson Bay Exploration and Development Company Ltd. completed horizontal loop EM surveys on a series of grids on the north side and to the southeast of Fisher Bay (SEM Assessment Files 63L11-0002, 63L14-0018). Five of the conductors on the north side of Fisher Bay were subsequently tested with diamond drill holes in 1974 and 1976 (SEM Assessment Files 63L-0014, -0017,-0018). Hole Des 111, which was collared in the highly tectonized and migmatized supracrustal sequence approximately one kilometre north of Fisher Bay, intersected banded hornblende-biotite-quartz granulite between 30.48 and 53.37 m containing several pyrrhotite-bearing siliceous intervals. One of these, between 46.45 and 46.60 m, yielded 75 percent pyrrhotite and assayed 0.7 g/t (0.02 oz/ton) Au and 6.2 g/t (0.18 oz/ton) Ag. Some of the other holes intersected minor amounts of Cu and Zn mineralization but contained no precious metals.

ii) **Jira Lake**: The Jira Lake showing, located approximately one kilometre east-southeast of the south end of Jira Lake (UTM co-ordinates 561120 m E, 6082390 m N), was discovered by D. Partridge in the spring of 1991. The occurrence lies near the southern contact of the Jira Lake Pluton (Padgham, 1966, 1968; Delaney, 1989), a leucocratic to locally mesocratic plagioclase porphyritic tonalite. Gold occurs in narrow, discontinuous, vuggy pyritic quartz veins within at least three narrow west-northwest-trending dextral shear zones that dip moderately to the north. In the most southerly of these shear zones, the only one well exposed, four quartz veins occur in 30 cm of sheared tonalite (Figure 2). Four grab samples from this zone contained 300 to 8700 ppb gold (Table 1).

b) Nickel, Chrome and Platinum Group Elements

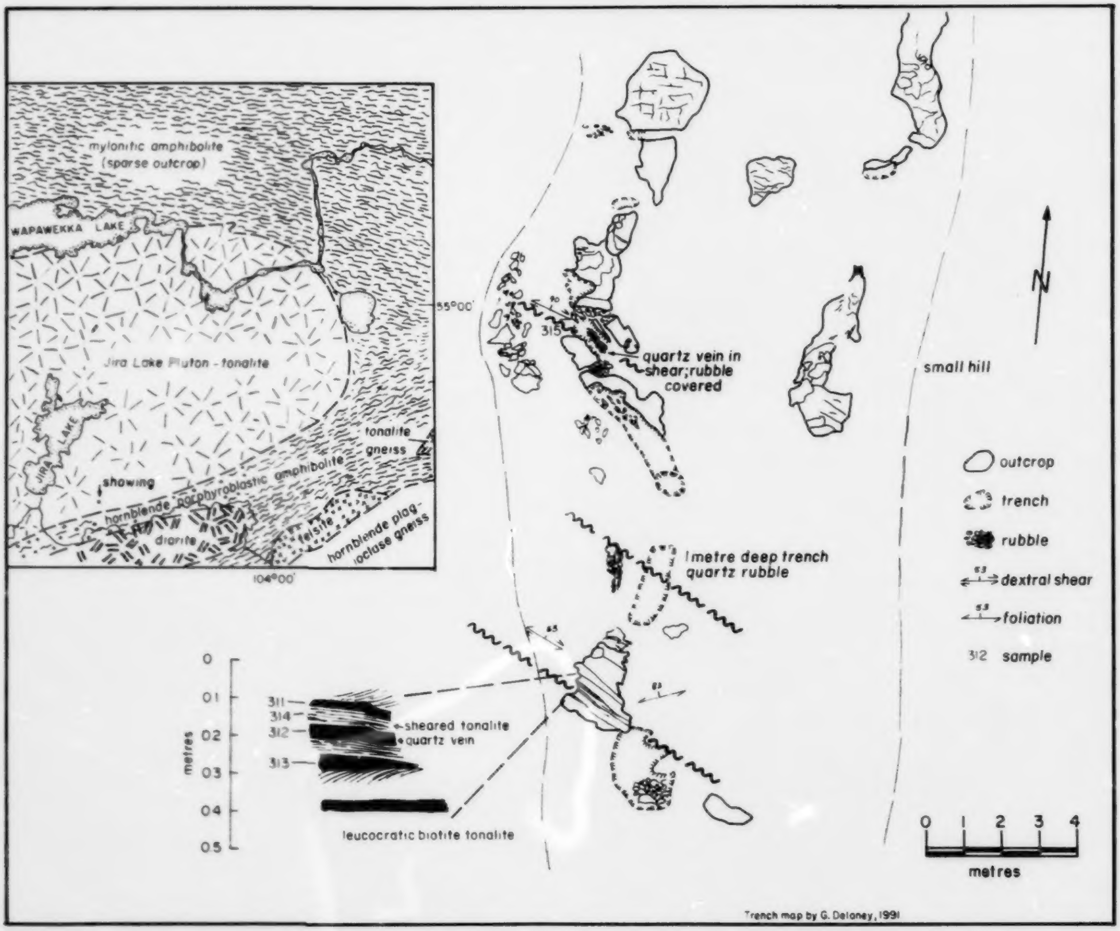
As the peridotite body on the north side of the entrance to the West Arm of Deschambault Lake was previously reported to have high nickel contents (Padgham, 1968), a suite of spaced samples was collected along the long axes of the intrusion. Samples were analyzed for 30 common elements as well as gold and platinum group elements (Table 2). These analyses not only confirm the high nickel content but also reveal a high concentration of chromium in the peridotite.

7. Recommendations for Prospecting

1) The discovery of shear-zone hosted gold mineralization near the southern margin of the Jira Lake Pluton invites a thorough evaluation of the rest of this pluton for gold mineralization. This search might also be extended to the west to cover adjacent plutons of a similar age that flank supracrustal rocks along Wapawekka Lake. These include the Radar Bay and Fisher Narrows Plutons on the south side of Wapawekka Lake and the Folkerson Lake Batholith and Lacey Lake Stock on the north side of the lake (Figure 1; Padgham, 1966, 1967; Delaney, 1989). Particularly favorable areas appear to be near the margins of the intrusions.

2) The Oskikebuk Narrows Pluton also warrants a thorough evaluation for gold mineralization. In the Brownell Lake area, the Brownell Lake Pluton and the Maynard Creek Pluton, which are similar to the Oskikebuk Narrows Pluton, contain gold mineralization. In particular, the microcline-porphyritic phase of the Brownell Lake Pluton, which appears to be the same as that in the Oskikebuk Narrows Pluton, hosts several gold occurrences.

3) Another target for gold exploration is high strain zones in supracrustal rocks of the Oskikebuk Lake Group. Because of generally poor exposure, identification of favorable structures will be hampered. Identification of structures can be achieved by completing close-spaced airborne electromagnetic and magnetic surveys and then using image processing techniques to enhance the data. The successful application of these techniques has recently been demonstrated in the Brownell Lake area (Delaney *et al.*, 1991).


mylonitic amphibolite (sparse outcrop)
WAPAWEKKA LAKE
55°00'
Jira Lake Pluton - tonalite
JIRA LAKE
showing
tonalite gneiss
hornblende porphyroblastic amphibolite
diorite
felsite
hornblende plagioclase gneiss
104°00'
N
quartz vein in shear; rubble covered
small hill
315
outcrop
trench
rubble
dextral shear
foliation
312 sample
1 metre deep trench quartz rubble
0
01
02
03
04
05
metres
311
314
312
313
sheared tonalite
quartz vein
leucocratic biotite tonalite
0
1
2
3
4
metres

Trench map by G. Delaney, 1991

Figure 2 - Outcrop and trench map Jira Lake showing. Inset is a geological sketch map of the Jira Lake area.

Table 1 - Geochemical Analyses for a Suite of Grab Samples from the Jira Lake Showing. Sample locations are shown on Figure 2. Data for all elements except gold were obtained by ICP analysis of a 0.500 gram sample digested with 10NL $HClO_3$-HNO_3-HF at 200°C and diluted to 10 ml with diluted Aqua Regia. This leach is partial for magnetite, chromite, barite, oxides of Al, Zr, and Mn and massive sulphide samples. Au analysis is by Atomic Absorption from a 10 g sample. Analyses by Acme Laboratories.

Element Samples	Mo ppm	Cu ppm	Pb ppm	Zn ppm	Ag ppm	Fe %	As ppm	Ba ppm	K %	W ppm	Zr ppm	Sn ppm	Sc ppm	Au* ppb
9122-311	3	160	3	57	0.5	1.71	5	272	1.68	9	21	2	4.4	1630
9122-312	2	108	2	35	0.2	1.11	2	59	0.36	6	3	6	1.1	650
9122-313	4	425	2	31	4.5	1.77	3	342	1.81	9	28	1	4.4	8700
9122-314	3	147	2	88	0.1	2.81	6	480	2.82	12	34	1	10.2	300
9122-315	3	284	4	34	0.4	2.84	2	50	0.28	4	6	7	1.3	480

Table 2 - Geochemical Analyses for a Suite of Grab Samples from Serpentinized Peridotite on the Peninsula on the Northeast Side to the Entrance to the West Arm of Deschambault Lake. Sample locations given as UTM coordinates. Data for all elements except Au, Pt, Pd, and Rh were obtained by ICP analysis of a 0.500 gram sample digested with 10NL $HClO_3$-HNO_3-HF at 200°C and diluted to 10 ml with diluted Aqua Regia. This leach is partial for magnetite, chromite, barite, oxides of Al, Zr, and Mn and massive sulphide samples. Au, Pt, Pd and Rh analysis is by ICP of a 10 g sample fused in a graphite furnace. Analyses by Acme Laboratories.

Element Samples	Cu ppm	Ni ppm	Co ppm	Mn ppm	Fe %	Cd ppm	V ppm	Ca %	P %	La ppm	Cr ppm	Mg %	K %	Sc ppm	Au* ppb	Pt ppb	Pd ppb	Rh ppb
9122-1192	120	409	55	1282	5.3	0.5	238	13.43	0.009	2	1475	8.35	0.2	50.8	13	21	18	1
9122-1193	84	536	68	1098	5.46	0.2	189	12.28	0.015	2	2086	9.52	0.02	48.2	19	31	42	1
9122-1194	1	1825	157	1181	8.29	0.5	72	2.79	0.014	2	3479	15.22	0.02	14.5	5	1	2	1
9122-1195	6	2504	175	1011	9.42	0.6	35	0.25	0.015	2	1717	18.32	0.01	8.9	5	20	4	1
9122-1196	93	1832	146	1181	8.73	0.4	65	2.18	0.009	2	1746	16.08	0.01	18.6	2	10	11	1

8. References

Bickford, M.E., Van Schmus, W.R., Macdonald, R., Lewry, J.F. and Pearson, J.G. (1986): U-Pb zircon geochronology project for the Trans-Hudson Orogen: Current sampling and recent results; *in* Summary of Investigations, 1986, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 86-4, p101-107.

Chiarenzelli, J.R., Lewry, J.F. and Landon, M. (1987): Bedrock geology, Iskwatikan Lake area: Evidence of Hudsonian juxtaposition of Proterozoic and Archean rocks along a ductile detachment surface; *in* Summary of Investigations, 1987, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 87-4, p46-51.

Chiarenzelli, J.R. (1989): The Nistowiak and Guncoat Gneisses: implications for the tectonics of the Glennie and La Ronge Domains, northern Saskatchewan, Canada; unpubl. Ph.D. Thesis, Univ. Kansas, 229p.

Delaney, G.D. (1988): Bedrock geological mapping, Brownell Lake area (part of NTS 63M-4 and 63L-13); *in* Summary of Investigations 1988, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 88-4, p8-19.

__________ (1989): Bedrock geological mapping, Maynard-Wapawekka Lakes area (part of NTS 73I-16, 63L-13, 63M-3 and 73P-1); *in* Summary of Investigations 1989, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 89-4, p17-27.

__________ (1990a): Structural hosted gold, Glennie Lake Domain, Lower Proterozoic Trans-Hudson Orogen; *in* Minerals Colloquium, January 1990, Ottawa, Geol. Surv. Can., Prog. Abstr., p18.

__________ (1990b): Gold mineralization, Brownell-Wapawekka Lakes Greenstone Belt, Trans-Hudson Orogen, Saskatchewan; Joint Annual Meeting, May 1990, Vancouver, Prog. Abstr., v15, pA31.

__________ (1991): Gold in the Glennie Domain, Lower Proterozoic Trans-Hudson Orogen, northern Saskatchewan; The Geological Society of CIM, First Annual Field Conference, Saskatoon, Sept. 1991, Prog. Abstr., p11.

__________ (in press): Gold in the Glennie Domain; Sask. Energy Mines, Misc. Rep.

Delaney, G.D., Carr, S.D. and Parrish, R.R. (1988): Two U-Pb zircon ages from the eastern Glennie Lake Domain, Trans-Hudson Orogen, Saskatchewan; *in* Radiogenic Age and Isotopic Studies, Geol. Surv. Can., Pap. 88-2, p51-58.

Delaney, G.D., Czornobay, B. and Earle, S. (1991): Utilization of image processed aeromagnetic and Landsat thematic image data for the improved understanding of the lithologic and structural characteristics of the Glennie Domain; The Geological Society of CIM, First Annual Field Conference, Saskatoon, Sept. 1991, Prog. Abstr., p11.

Delaney, G.D., Heaman, L.M., Kamo, S., Parrish, R.R., Slimmon, W.L. and Reilly, B.A. (1990): U-Pb sphene/zircon geochronological investigations; *in* Summary of Investigations 1990, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 90-4, p54-57.

DeLury, J.S. (1926): Wapawekka and Deschambault Lakes area, Saskatchewan; Geol. Surv. Can., Summ. Rept. 1924, Pt. B. p23B-50B.

Hornbrook, E.H.W., Garrett, R.G., Lynch, J.J. and Beck, L.S. (1975): Regional lake sediment reconnaissance data, east-central Saskatchewan; Geol. Surv. Can., Open File 266, 108p.

__________ (1977): Regional lake sediment reconnaissance data, east-central Saskatchewan; Geol. Surv. Can., Open File 488.

Hornbrook, E.H.W., Lynch, J.J., Friske, P.W.B., Lund, N.D. and Schmitt, H.R. (1985): Regional lake sediment reconnaissance data, east-central Saskatchewan; Geol. Surv. Can., Open File 1129, 145p (8 maps).

Langford, F.F. (1973): The geology of the Wapawekka area, Saskatchewan (73I); Sask. Dep. Miner. Resour., Rep. 147, 36p.

Lewry, J.F. (1981): Lower Proterozoic arc-microcontinent collisional tectonics in the western Churchill Province; Nature, v294, no. 5836, p69-72.

Lewry, J.F. and Collerson, K.D. (1990): The Trans-Hudson Orogen: Extent, subdivision and problems; *in* Lewry, J.F. and Stauffer, M.R., (eds.), The Early Proterozoic Trans-Hudson Orogen of North America, Geol. Assoc. Can., Spec. Pap. 37, p75-94.

Lewry, J.F. and Macdonald, R. (1988): Observations on deformation in the Glennie Domain and Hanson Lake Block; *in* Summary of Investigations 1988, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 88-4, p35-41.

Lewry, J.F. and Slimmon, W.L. (1985): Compilation bedrock geology, Lac La Ronge, NTS area 73P/73I; Sask. Energy Mines, Report 225 (1:250 000 scale map with marginal notes).

Lewry, J.F., Thomas, D.J., Macdonald, R. and Chiarenzelli, J. (1990): Structural relations in accreted terranes of the Trans-Hudson Orogen, Saskatchewan: telescoping in a collisional regime; *in* Lewry, J.F. and Stauffer, M.R. (eds.), The Early Proterozoic Trans-Hudson Orogen of North America, Geol. Assoc. Can., Spec. Pap. 37, p75-94.

Macdonald, R. (1975): Compilation geology, Pelican Narrows (63M) and Amisk Lake (63L) areas; *in* Summary of Investigations 1975 by the Saskatchewan Geological Survey, Sask. Dep. Miner. Resour. p44-47.

__________ (1981): Compilation bedrock geology: Pelican Narrows and Amisk Lake areas; 1:250 000 scale prelim. map with Summary of Investigations 1981, Saskatchewan Geological Survey, Sask. Miner. Resour., Misc. Rep. 81-4.

__________ (1987): Update on the Precambrian geology and Domainal classification of northern Saskatchewan; *in* Summary of Investigations 1987, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 87-4, p87-104.

McInnes, W. (1913): The basins of the Nelson and Churchill Rivers; Geol. Surv. Can., Mem. 30.

Padgham, W.A. (1966): The geology of the Wapawekka Narrows area (north half), Saskatchewan (73I-16-N); Sask. Dep. Miner. Resour., Rep. 87, 113p.

__________ (1967): The geology of the Wapawekka Lake area (north half), Saskatchewan (73I-15-N); Sask. Dep. Miner. Resour., Rep. 103, 57p.

__________ (1968): The geology of the Deschambault Lake district comprising the Oskikebuk Lake area, Viney Lake area (west half), the northern part of the Ballantyne Bay area, and the northeast quarter of the Limestone Lake area, Saskatchewan (63-L-13, 63-L-14-W and parts of 63-L-11, -12); Sask. Dep. Miner. Resour., Rep. 114, 92p.

Slimmon, W.L. (1988): Bedrock geological mapping, Gee Lake area (part of NTS 63M-3 and -4); *in* Summary of Investigations 1988, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 88-4, p26-34.

__________ (1989): Bedrock geological mapping: Sadler Lake area (part of NTS 63M-4 and -5); *in* Summary of Investigations 1989, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 89-4, p28-33.

Thomas, D.J. (1988): Bedrock geological mapping, Palf Lake area (part of NTS 63M-3 and -4 and 63L-13 and -14); *in* Summary of Investigations 1988, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 88-4, p20-25.

Revision Bedrock Geological Remapping of the Southeast Arm, Deschambault Lake-Northern Lights Area (Parts of NTS 63L-10, -11 and -14)[1]

J.F. Lewry[2], T.I.I. Sibbald and N. Hayward[3]

Lewry, J.F., Sibbald, T.I.I. and Hayward, N. (1991): Revision bedrock geological remapping of the Southeast Arm, Deschambault Lake-Northern Lights area (parts of NTS 63L-10, -11 and -14); *in* Summary of Investigations 1991, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 91-4.

This summer, the authors collaborated in 1:20 000 scale re-mapping between the Southeast Arm of Deschambault Lake and Tulabi Brook (Figure 1). The area, previously mapped by Padgham (1968) and Macdonald and Posehn (1976), lies in the southeastern Glennie Domain and straddles a fault array interpreted as part of the Tabbernor fold-fault zone. Lewry also examined lakeshore outcrops in the Twin Bays area, Deschambault Lake, northwest of the map area. The main objectives were to:

a) re-examine the most southerly exposed part of the Tabbernor zone, define the most important related structural discontinuities and establish the degree of continuity, or otherwise, of older lithostructural elements between the Glennie Domain and Hanson Lake Block;

b) explore possible eastern extensions of early high-strain zones already defined in the southern Glennie Domain, mainly the Hartley Lake shear zone (Delaney, 1988), and establish their relationship, if any, both to the Tabbernor zone and the 'Pelican Slide' in the Hanson Lake Block (Lewry and Macdonald, 1988; Lewry *et al.*, 1989; Lewry, 1990); and

c) re-map the Northern Lights volcanics (Macdonald and Posehn, 1976), thought to be of similar age to volcanic rocks in the Hanson Lake area, and evaluate their economic potential.

1. Description of Rock Types

Macdonald and Posehn (1976) divided the area into several more or less distinctive fault-bounded segments; the Southeast Arm segment west of Sarginson Lake, the Unser Lake schist belt passing through Sarginson Lake, the Northern Lights volcanic segment east of this lake; and the Tulabi Lake segment (which we did not map) east of the Tulabi Brook fault. We use these names in the current report.

a) Unit 1 - Mafic Volcanics

These rocks, forming part of the Northern Lights segment, occupy the core of a major antiform east of Sarginson Lake between the old and new highways, and are sandwiched by mixed volcaniclastics of unit 2. As noted by Padgham (1968) and Macdonald and Posehn (1976), they retain primary volcanic features such as pillows, flow-top breccias and amygdales, especially in the antiform hinge zone where dominantly linear, steeply north-plunging fabrics produce apparently little-deformed cross-sections on subhorizontal outcrop surfaces. These are well displayed in borrow pits north of the old highway.

The mafic volcanics typically comprise massive units several metres thick. They are dark green, inequigranular, variably foliated and lineated amphibolites with subidioblastic-xenoblastic hornblende porphyroblasts (20 to 30 percent) up to 0.5 cm diameter set in a fine- to medium-grained amphibole-feldspar matrix. Pale yellow-green epidosite 'bombs' up to several tens of centimetres in diameter occur as irregularly shaped replacements. Well-laminated mafic tuffs, displaying primary grading which indicates southward younging, are interlayered with more massive, in part pillowed flow units, north of the old highway. Blue quartz-eye ± feldspar-phyric felsic 'tuffs' are interlayered with the lower part of the mafic volcanics, notably in the area directly east of Sarginson Lake.

b) Unit 2 - Mixed Volcaniclastics

Mixed volcaniclastic rocks form the dominant unit in the Northern Lights segment. As noted by Padgham (1968, unit 2), they are extremely variable in character, but such variation is not on a scale that allowed subdivision into mappable units in the present study. Compositionally, the rocks range from mafic to felsic, with intermediate to felsic types predominating. Layering is variably well-developed on scales from millimetres to tens of metres: most typically it is on a relatively large (metre) scale and cryptic. Unequivocal primary volcanic features were not

---

(1) Saskatchewan Project A.147, from which this report derives, was funded in 1991 under the Canada-Saskatchewan Partnership Agreement on Mineral Development 1990-95.

(2) Department of Geology, University of Regina, Regina, Saskatchewan
(3) University College of Wales, Aberystwyth, U.K.

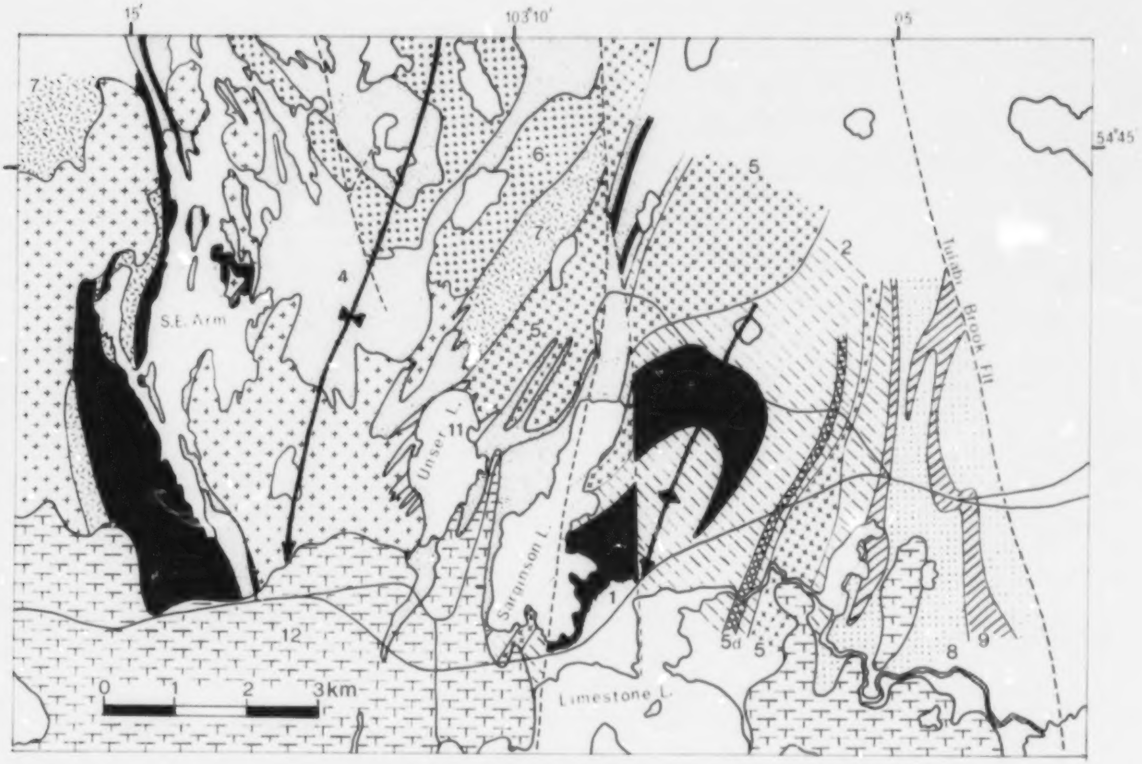


Figure 1 - General geology of the Southeast Arm, Deschambault Lake-Northern Lights area. Indicated in black are mafic metavolcanics (unit 1) east of Sarginson Lake, and mafic gneisses (unit 3) in the Southeast Arm area. Other units, indicated by number and pattern, are: 2 (oblique dash) = mixed volcaniclastics; 4 (fine stipple) = psammitic to psammopelitic metasediments; 5 (crosses) = early synkinematic tonalite to granodiorite; 5d (heavy oblique line) = diorite; 6 (heavy dots) = early megacrystic pink granodiorite/granite; 7 (random dash) = early equigranular pink granite/granodiorite; 8 (regular fine dot) pink and grey quartzofeldspathic granulites; 9 (oblique lines) = mafic granulites; 10 (upright cross) = late-kinematic leucogranodiorite, aplogranite and pegmatite (Jan Lake suite); 11 (in Unser Lake basin mainly) = complex late muscovite-tourmaline-beryl pegmatites; 12 (in the south) = Ordovician dolomites and basal sands. Synformal fold axial surface trace = Unser Lake fold; Antiformal fold axial surface trace = Sarginson Lake antiform.

observed, although Macdonald and Posehn (1976) suggested the presence of flows and tuffaceous units.

Dark green, amphibole-rich (>30 percent) mafic end members are transitional to intermediate types characterized by amphibole ± biotite ± garnet (<30 to >15 percent) and these in turn grade into grey, biotite ± garnet ± amphibole ± magnetite (<15 percent) bearing felsic members. All are uniformly medium to fine grained, typically moderately to strongly foliated and in places submylonitic, although they exhibit granoblastic textures indicating post-tectonic recrystallisation. Hornblende commonly forms porphyroblasts up to 0.5 cm diameter in mafic units. Garnets, averaging 1 to 3 mm in diameter but locally up to 1 cm, and feldspar megacrysts up to 0.5 cm diameter are common in felsic units but unusual in mafic rocks. 'Quartz eyes' up to a few millimetres in diameter, typically iridescent blue, occur in the most felsic units, particularly those occurring along the northern contact of the mixed volcaniclastics and mafic volcanics. Quartz-eye felsic rocks are also interlayered with mafic volcanics in this region. Irregularly shaped epidosite lenses and replacements are commonly present in mafic to intermediate layers. Fine-grained 'cherty' siliceous layers and sulphidic gossanous zones occur sporadically throughout the volcaniclastics, the former particularly in well laminated parts of the assemblage. Pink quartz-K feldspar-biotite 'granite' pegmatites are common in parts of the mixed volcaniclastic unit, particularly east and north of the old Hanson Lake road. Most appear to be deformed, although fabrics are less well defined than in the host volcaniclastics.

We interpret rocks in this unit as predominantly clastic reworked and possibly primary volcanics. They were deposited subaqueously in a relatively distal setting, because coarse fragmental units are rare.

A single outcrop area (100 x 50 m) comprising massive, medium- to coarse-grained amphibolite (unit 2a) occurs north of the old Hanson Lake road 1 km northwest of its eastern junction with the new highway. The extreme mafic composition of this rock (ca. 100 percent amphibole) and its homogeneous character suggest that it may be a small ultramafic intrusion.

c) Unit 3 - Mafic Gneisses and Schists

Hornblende-rich schistose gneisses underlie much of the area westward from the southwestern shore of Southeast Arm and form thinner belts within dominant metasediments extending through the latter.

Compositionally, much of the larger area of exposed mafic rocks (unit 3a) is broadly homogeneous, but there is marked heterogeneity locally. The dominant component, a medium-grained amphibolite locally grading to more plagioclase-rich hornblende gneiss, is strongly foliated and lineated. In places it is poorly- to well-laminated; however, much of the lamination appears to be strain-induced rather than compositional. Pervasively developed idioblastic to subidioblastic hornblende porphyroblasts, from a few millimeters to 2 cm diameter, are typical of most outcrops. Generally, they appear to overgrow early kinematic fabrics (S_1/L_1), but local augening suggests later reactivation of the main foliation. Semi-continuous to lensoid interbanding of richly porphyroblastic and non-porphyroblastic amphibolite is variably developed. Greenish epidosite streaks and lenses, and chocolate-brown to cream coloured, carbonate-rich, in part diopsidic lenses and discontinuous bands occur in many outcrops. It is unclear whether the latter are disrupted primary layers or tectonometamorphic-hydrothermal(?) segregations derived from the host protolith. Elsewhere, thin, fine-grained, in part amphibolite layers and light-grey felsic units up to a metre or more thick may represent either primary volcanic layers or thin hypabyssal sheets. Indefinite to distinct compositional banding, possibly representing primary supracrustal layering, was seen in a few clean outcrops but no totally unequivocal primary features were noted.

The protolith of this mafic subunit is uncertain. Macdonald and Posehn (1976) provisionally interpreted it as highly strained to mylonitic metagabbro. Although much of the unit is compositionally relatively homogeneous and lacks definite primary layering, no coarser-grained, less-strained, unequivocal gabbroic protolith relicts were observed and the abundant large hornblendes typifying the unit are clearly the product of metamorphic blastesis rather than primary igneous relicts. Moreover, local compositional layering suggests that at least parts of the unit are supracrustal in origin. Provisionally, we consider unit 3a to comprise highly-strained mafic to intermediate metavolcanics, possibly even the lateral equivalents of unit 1.

The thinner mafic belts (unit 3b) comprise extremely fine-grained, intensely foliated amphibolitic schists which, although finely laminated, are rather homogeneous overall. Ill-defined, possibly transposed layering was noted in only two localities. These rocks may represent highly strained mafic metavolcanics but their overall uniformity and the absence of any gradation into, the enclosing metasediments (unit 4), suggests that they are more likely to be strongly deformed hypabyssal intrusive sheets.

d) Unit 4 - Psammitic to Psammopelitic Metasediments

Clastic metasediments, mainly medium- to fine-grained psammitic to psammopelitic metagreywackes variably injected by late granitoid phases, underlie much of the Southeast Arm segment. Psammitic variants are generally mid- to light-grey, medium-grained, equigranular, granoblastic rocks with up to 10 percent biotite, variable quartz and plagioclase and generally low K-feldspar content. Biotite is typically dispersed throughout the rock but also defines schistose partings. Less commonly, it also occurs as scattered porphyroblastic flakes of somewhat larger grain-size than quartz and feldspar. The psammopelites are broadly similar but have more biotite, as large individual flakes and prominent lensoid multicrystalline folia. Minor pelitic components generally occur as thin laminae in the psammite-psammopelite assemblage. Compositional banding, probably representing transposed primary layering, occurs on all scales from millimetres to metres; psammitic parts of the unit tend to be more massively bedded.

Andalusite porphyroblasts up to 3 cm across are locally abundant in pelitic-psammopelitic layers throughout the area. A few outcrops, mainly on islands in Southeast Arm, include layers with plentiful small lensoid sillimanite- or sillimanite-quartz-muscovite faserkiesel.

The metasediments are generally injected by abundant, pervasive quartz veins from 1 mm to a few centimetres thick, which may comprise 5 percent or more of outcrops. Most veins are concordant to early tectonic foliation, isoclinally folded and severely boudinaged. Around northern parts of Southeast Arm and the area immediately to the east, early neosomal veins commonly include abundant K-feldspar, rather than simply quartz. Later, generally larger, discordant, massive to foliated quartz veins also occur locally.

As noted, most metasediments are medium- to fine-grained granoblastic rocks; however, in the Unser Lake schist belt, between Sarginson and Unser Lakes, they are generally extremely fine-grained to aphanitic, and dark grey to black. Similar very fine-grained variants also form ill-defined zones extending along Southeast Arm, within more 'normal' metasediments. Despite their finer grain-size, they display abundant early quartz veining, local transposed layering and andalusite-bearing units, and appear to be psammitic to psammopelitic. Some shoreline outcrops show evident strain gradation from 'normal' medium-grained metasediments into intensely deformed, very fine-grained zones in which relict compositional layering becomes progressively less evident and subvertical lineation intensifies. Thus, we interpret such fine grain-size as a function of early ductile high-strain and dynamic recrystallisation, rather than of primary lithological variation. Such rocks are essentially mylonitic in character.

In two outcrops adjacent to the old highway east of Unser Lake, the metasediments incorporate units of very fine-grained to aphanitic, streakily laminated, cream to pink leucocratic 'felsite' (unit 4a) with an intense subvertical stretching lineation. It is unclear whether these are highly strained felsic volcanics or mylonitized granitoid sheets.

e) Early Granitic Rocks

Three distinct plutonic rock types were emplaced prior to or during the earliest documented deformation event in the area.

Unit 5 - Grey Tonalite-Granodiorite

A large tonalite-granodiorite body (unit 5a) extends northward from Unser Lake across the northeast part of the Southeast Arm segment. Though compositionally homogeneous, it displays great heterogeneity of early ductile kinematic fabric, ranging from weakly foliated to mylonitic in character, with locally prominent lineation. Weakly strained variants are medium to coarse grained, non-porphyritic, grey-white both on weathered and fresh surfaces, contain both biotite and minor hornblende (total mafics < 10 to 15 percent), plagioclase, quartz and apparently minor K-feldspar. Protomylonites to mylonites are progressively finer-grained, darker grey rocks with a gneissose to finely laminated fabric. Much of the body possesses a variably well-developed protomylonitic C-S kinematic fabric. Mylonite-ultramylonite zones up to a metre or more wide are locally abundant.

Another major body of generally grey to pink weathering, homogeneous, weakly megacrystic biotite granodiorite (unit 5b) lies north and east of Sarginson Lake, and extends southwards as a narrow tongue along the northeast shore of this lake. Outcrops on a point at the southern end of Sarginson Lake probably represent a fault-displaced extension of the same body. The rock has a medium-grained groundmass of quartz (20 to 30 percent), biotite (5 to 10 percent) and accessory magnetite enclosing somewhat coarser-grained, subhedral to rounded plagioclase megacrysts averaging 0.5 cm in diameter but locally up to 1.0 cm. Kinematic fabrics are dominantly linear throughout most of the body and are variably weakly to strongly developed. Outcrops adjacent to the northeast shore of Sarginson Lake display an intense stretching lineation. Contacts are generally unexposed; where seen, they are indistinct (particularly along the northeast side of Sarginson Lake) due to apparent marginal injection, rafting of adjacent host rocks and subsequent strain transposition. Xenolithic material is otherwise typically absent. Pink pegmatite, granite and aplite form weakly deformed crosscutting sheets and diffuse zones within the body north of the old highway and seem to become more widespread northwards.

Grey to locally pink (hematitic), equigranular medium-grained (2 to 5 mm), homogeneous biotite granodiorite (unit 5c) forms a well-exposed mappable concordant unit extending northward from Grassberry River. Typically, it is strongly foliated but clearly derives from a coarse grained plutonic protolith. Deformation and dynamic recrystallization have resulted in streaking and replacement of original large quartz, feldspar and biotite crystals by annealed multicrystalline aggregates.

A thin, sheet-like body of quartz dioritic to dioritic composition (unit 5d) occurs in the mixed volcaniclastic rocks west of the unit 5c granodiorite. It also is medium grained (1 to 2 mm), strongly foliated, equigranular and derived from a coarser grained protolith, as indicated by streaked aggregates of hornblende and feldspar. The rock mainly comprises plagioclase and hornblende (20 to 30 percent), but quartz is evident in some places.

Unit 6 - Pink Megacrystic Granodiorite-Granite

This rock outcrops as somewhat irregular, apparently folded bodies in the northern part of the Southeast Arm segment and also forms a subordinate part of early plutons west of Southeast Arm. The rock is generally pink to grey-pink, medium grained, inequigranular with sparse to abundant, 1 to 2 cm diameter K-feldspar augen. It is generally strongly foliated, locally also strongly lineated, and is generally more homogeneously deformed than unit 5a, although there are local narrow mylonite zones.

Unit 7 - Equigranular Pink Granite-Granodiorite

This unit forms a discrete, elongate mass lying between tonalite-granodiorite (unit 5a) and megacrystic granodiorite-granite (unit 6) north of Unser Lake, and also comprises most of the early plutons west of Southeast Arm. The rocks are pink to grey-pink, medium grained and non-megacrystic, but there is local gradation to sparsely megacrystic variants. The body north of Unser Lake is generally uniformly strongly foliated. Fabrics in the area west of Southeast Arm range from L-S tectonites with dominant planar fabric to almost pure L-tectonites.

f) Rocks of Uncertain Origin

The easternmost part of the Northern Lights segment, directly west of the Tulabi Brook fault, is underlain by compositionally bimodal quartzofeldspathic and hornblendic granulites (term used in a textural sense) of uncertain protoliths. Pink pegmatite sheets cut both the quartzofeldspathic and hornblendic granulites.

Unit 8 - Quartzofeldspathic Granulites

Macdonald and Posehn (1976) subdivided these rocks into pink granitoids and felsic volcanics. Although we recognized pink and grey quartzofeldspathic granulites as viable subdivisions, they are not wholly coincident with Macdonald and Posehns' units. Moreover, contacts between the two are indistinct, somewhat subjective and can be arbitrarily located: thus they are not differentiated by discrete boundaries on the accompanying map.

Pink quartzofeldspathic granulites (8a) are homogeneous, medium-grained (1 to 2 mm), equigranular-granoblastic rocks comprising quartz, feldspar, biotite (5 to 10 percent) and minor magnetite. They are typically moderately to strongly foliated and lineated. Possible xenoliths up to a few tens of centimetres long were observed in two localities.

Grey quartzofeldspathic granulites (8b) are also medium-grained (1 to 3 mm), equigranular-granoblastic, weakly to strongly foliated and lineated rocks composed mainly of quartz, feldspars and biotite (0 to 15 percent). Red garnet, magnetite and, uncommonly, hornblende are minor constituents. Compositional variation is seldom apparent at outcrop scale but in a few places there is an indistinct layering defined by variations in quartz-feldspar ratio (quartz content may reach 50 percent) or concentration of minor constituents (e.g. garnet). Rarely, thin units of hornblende granulite are interlayered with otherwise homogeneous grey quartzofeldspathic granulite.

The protoliths of the quartzofeldspathic granulites are uncertain. Both Padgham (1968) and Macdonald and Posehn (1976) interpreted parts of these units as igneous and others as supracrustal rocks. Macdonald and Posehn suggested that their pink granitoids (unit Gl) might represent "in part layered supracrustals, locally remobilized", an interpretation which is permissible on the basis of observed field relationships. However, their interpretation of other quartzofeldspathic granulites as felsic volcanics is not supported by recognizable primary volcanic features. It is considered more likely that precursors were psammitic or arkosic metasediments. In view of the evident high strain, however, another possibility is that both sub-types are mylonitized granitoids.

Unit 9 - Hornblende Granulites

These form units within the quartzofeldspathic granulites from a few metres to several hundred metres thick. They are pale green, medium grained (1 to 2 mm), essentially equigranular-granoblastic, moderately to strongly foliated, poorly to well layered and, on average, of mafic to submafic (±30 percent mafic content) composition. Layering, on a millimetre to metre scale, is defined by variations in the ratio of mafic to felsic constituents. The most mafic layers comprise dominant hornblende and feldspar, whereas felsic end members are quartz±feldspar±hornblende±biotite±garnet±magnetite assemblages. Garnets are commonly weakly porphyroblastic and up to 3 mm in diameter. In a few places, feldspar megacrysts of similar size occur in layers of felsic-intermediate composition.

Unit 10 - Late Granites

Late- to post-kinematic leucogranite to leucogranodiorite has extensively invaded both supracrustal rocks and early plutonic units throughout the Southeast Arm segment. These late intrusive phases mainly comprise medium-grained, homogeneous, non-megacrystic grey-white 'leucogranodiorite' to red-pink aplogranite which locally grades into, or is cut by coarser-grained to pegmatitic leucogranite. Mafic content is generally less than 5 percent. Biotite, the dominant mafic phase, occurs as small uniformly dispersed flakes; small red garnets are a common minor phase and magnetite is also present locally. The relationship between white leucogranodiorite and pink aplogranite is variable and equivocal. Locally, aplogranite clearly intrudes the leucogranodiorite; elsewhere, one appears to grade insensibly into the other across outcrop and it is unclear either from outcrop or hand-specimen if they represent distinct intrusive phases, or if the colour change is a function of gradational K-metasomatism or even of differential weathering. Typically, the leucogranodiorite displays a weak linear (less commonly planar) tectonic fabric, whereas the pink leucogranite phase more commonly appears massive. However, this is not consistent; many leucogranodiorite outcrops are apparently massive and some aplogranites have a weak fabric.

Such rocks, together with variable amounts of associated, apparently broadly coeval pegmatite, form major intrusive masses, particularly west of Southeast Arm and in the west-central part of the area between Southeast Arm and Unser Lake. Metasediments west and north of Unser Lake and along Southeast Arm are commonly pervasively, irregularly intruded by late granitic bodies ranging in size from sheets less than a metre thick, to mappable sheetlike to irregular masses. Older granodiorites (units 6 and 7) west of Southeast Arm and in the northern part of the map area are also irregularly, complexly invaded by late leucocratic phases, as is the northern part of unit 5b. Such material is generally absent in the Unser Lake schist belt and Northern Lights segment.

Late leucogranodiorites and aplogranites of unit 9 are generally identical in range of petrographic character, field relations and fabric variation to rocks of the 'Jan Lake suite' in the Jan-Tulabi Lakes area, Hanson Lake Domain (Lewry, 1990). Farther east they are thought to be approximately coeval with them.

Unit 11 - Pegmatites

Pegmatites of several ages, differing mineralogy and/or degree of deformation occur in the area.

Mineralogically simple (feldspar-quartz±biotite), red-pink granite pegmatite sheets up to 2 m thick, which occur commonly in the foliated megacrystic and non-megacrystic pink granodiorite-granite bodies (units 6 and 7), are probably co-magmatic with their hosts. They are folded and/or boudinaged and possess a moderate to strong axial planar foliation (S_1) parallel to and evidently coeval with that in the enclosing pluton. Less abundant, similarly deformed early pegmatite sheets occur in the tonalite-granodiorite (subunit 5a).

Younger, generally massive to weakly deformed, crosscutting sheets and irregular masses of mineralogically simple (quartz-K-feldspar-plagioclase±biotite) red-pink granite pegmatite, apparently coeval with or slightly younger than unit 10 aplogranite occur sporadically and irregularly throughout the area, cutting most other map units. In the Southeast Arm segment, they are most common in the metasediments, in association with aplogranite, and also cut the late grey-white leucogranodiorite. In the Northern Lights segment, pink pegmatites occur in most rock units, but are particularly strongly developed in the mixed volcaniclastics north of the old highway. These pegmatites are locally boudinaged and in places display a weak internal tectonic fabric.

Mineralogically complex, biotite-muscovite granite pegmatites, commonly containing beryl, tourmaline, garnet and possibly other rare element minerals (e.g. spodumene) (cf. MacDougall, this volume), are mainly exposed as large masses around the shoreline of, and islands in, Unser Lake but are also encountered north and west of this lake, cutting grey tonalite-granodiorite (unit 5a) or metasediments. In the Northern Lights segment, black tourmaline ± muscovite-bearing granite pegmatites were noted at two localities in volcaniclastic rocks outcropping along the eastern contact of the megacrystic granodiorite (subunit 5b) and at one locality north of the old highway.

Typically, the complex pegmatites are very coarse grained, locally showing well-shaped K-feldspar crystals up to 1 m across, and commonly contain 'leaf-like', subradiate muscovite-quartz intergrowths. Their size and morphology was not clearly established, since contacts are generally unexposed. However, inland occurrences north of Unser Lake suggest that they are probably large northeast- to north-northeast-trending sheets. They clearly postdate D_1 fabrics in the metasediments and grey tonalite-granodiorite, but their age relative to the D_2 event is uncertain, as locally, they display a weak to well-defined internal foliation which may be of D_2 age. Their age and genetic relation to the late granitoids of unit 10 is also uncertain.

2. Structure

a) Early Fabrics

Supracrustal and mafic gneisses are typically well foliated and locally mylonitic. In the metasediments, S_1 foliation is paralleled by transposed primary layering and early quartz or quartz/feldspar veining. In the mafic gneisses (unit 3) and some metasediments, it is accentuated by fine tectonometamorphic segregation lamination. Isoclinally F_1 folded and boudinaged quartz veins are ubiquitous in the metasediments and minor F_1 intrafolial folding of primary compositional layering in metasediments, mixed volcaniclastic and mafic gneisses is locally visible.

Early plutonic rocks (units 5, 6, 7) generally also display a well-developed early foliation presumed to be of the same age as S_1 foliation in supracrustal rocks. In units 5 and 6, planar fabrics range from weak to locally protomylonitic-orthomylonitic, typically with markedly heterogeneous strain on an outcrop scale. In unit 5a tonalite-granodiorite, this early foliation is commonly a well-defined compound C-S shear fabric.

A generally steeply-plunging to subvertical lineation is commonly, albeit sporadically, evident throughout the western part of the area, in mafic gneisses, metasediments (particularly finer-grained variants) and early granitoid units (6, 7), particularly along and to the west of Southeast Arm. It is seen only rarely in the tonalite-granodiorite body (unit 5a) north of Unser Lake. East of Sarginson Lake, similar lineation is prominent in the mafic metavolcanics and porphyritic granodiorite (unit 5b) and variably developed in other units. Its orientation is also variable throughout the Northern Lights segment. In some cases, particularly in outcrops of porphyritic granodiorite (unit 5b) and early pink granite (unit 7), it is clearly a stretching lineation defining an almost pure L-tectonite fabric. Elsewhere, it is generally part of a compound L-S_1 fabric.

b) Northerly Folding

Structurally, the area is dominated by the Unser Lake-Sarginson Lake fold pair. The Unser Lake fold is a southward-closing, subvertically-plunging F_2 major structure with a sub-vertical northerly-trending axial surface. It refolds both S_1 foliation and transposed primary layering. Steeply plunging, open to tight F_2 minor folds, with a sense of asymmetry consistent with their being coeval with the Unser Lake fold, are developed throughout the area, particularly in metasediments. Locally, they are accompanied by a weak to well-developed, spaced to pervasive axial plane schistosity (S_2). A weak to indefinite subvertical north- to northeast-striking foliation locally developed in the late granitic units is also tentatively interpreted as S_2. The Sarginson Lake fold is northward closing, antiformal and also steeply plunging. Minor structures, which are uncommon, have sub-vertical north-northeast-trending axial surfaces. Strong rodding fabrics, reflecting stretching parallel to the fold axis, occur in the fold core but axial plane fabrics are absent.

c) Age of Linear Fabric Elements

Hinge lines of F_2 minor folds, rare measurable F_1 folds, rodding defined by disrupted and boudinaged early quartz veins, mineral lineation, rodding (stretching) lineation in supracrustal and early granitoids, and a weak to indefinite lineation in late leucogranite-leucogranodiorite phases are essentially coaxial and for the most part steeply northerly to southerly plunging, parallel to the hinges of the major Unser Lake-Sarginson Lake folds. Despite this coaxiality, some of these linear fabric elements are probably of D_1 rather than D_2 age. For example, in the early granitoid units there is transition from rocks with almost pure L-tectonite fabrics into those with L-S or S fabrics. There is no indication that the L-fabric represents an S_1/S_2 intersection. Locally, there is also evident small-scale strain-partitioning between granitoid lenses with essentially L-fabrics and enclosing supracrustals with L-S to S fabrics, but no indication that this results from multiple deformation events. Subvertical quartz rodding in metasediments generally seems to be related to intrafolial F_1 folding of quartz veins and concomitant development of regional S_1 foliation.

On the basis of observed character of D_2 structures, it seems unlikely that D_2 extensional strain was sufficient to cause total transposition of such D_1 linear fabric elements into parallelism with D_2 fold hinges. Thus, we suggest that second phase linear fabrics, including the hinges of the major Unser Lake-Sarginson Lake fold pair, developed approximately parallel to the prior D_1 stretching direction.

d) Significance of Spatial Variation in D_1 Strain Intensity

There is a poorly delimited spatial variation in D_1 strain in the map area. The metasediments extending northeast of Unser Lake appear more deformed than those in the core of the Unser Lake fold to the west, and are categorized as protomylonites and mylonites. Deformation also increases, albeit irregularly, eastward across the intervening tonalite-granodiorite (unit 5a). Similarly, rocks in the package extending though Southeast Arm are generally more highly- and heterogeneously-strained than those immediately to the west and east. On this basis, we recognize a broad, indefinitely bounded belt of heterogeneous D_1 ductile high-strain extending through Southeast Arm, around the closure of the Unser Lake fold and up along the eastern limb of the latter, north of Sarginson Lake. This high-strain zone is preferentially concentrated in the metasediments, but also affects mafic gneisses and early granitoids. The high-strain fabric includes a well-developed, now generally subvertical L_1 stretching lineation.

To the east, deformation appears to be strongest in the mixed volcaniclastics forming the eastern limb of the Sarginson Lake antiform, particularly in a northerly-trending zone through Hand Lake.

e) Shear-Sense Indicators

Well-defined shear-sense indicators were widely observed in unit 5a and rarely in unit 6, but not in outcrop areas which also displayed a well-developed L_1 stretching lineation. Consequently, sense of displacement in the D_1 high-strain zone defined above was not determined. Locally well-developed C-S fabric relations observed on subhorizontal outcrop surfaces of the tonalite-granodiorite (unit 5a) northeast of Unser Lake record a consistent sinistral shear-sense but the three-dimensional significance of this is unclear. The C-S fabric is transected both by thin secondary shear bands and by apparently later, wider, discrete ductile mylonitic-ultramylonitic shears which display dextral shear-sense, opposite to that shown by the C-S fabric. Such displacement is consistent with flexural slip along the eastern limb of the D_2 Unser Lake fold and the secondary shear zones are thus tentatively interpreted as D_2 structure.

f) Late Faulting

Four significant late faults, together forming part of the Tabbernor Lake fault zone can be identified on the basis of current mapping:

a) The Tulabi Brook fault was identified by Padgham (1968, p.65) as "probably the most important fault in the Limestone Lake area". Outcrops along Tulabi Creek south of the Hanson Lake road are hematized, epidote stained, silicified and net-veined by quartz. Displacement on this structure is uncertain; Padgham (1968, p.66) favoured a dextral strike slip displacement of 1.5 km and an even larger component of vertical movement.

b) The West and East Sarginson Lake faults, also recognized by Padgham (1968) and Macdonald and Posehn (1976) run north-south through Sarginson Lake and are defined by unequivocal sinistral strike displacement of lithological units, principally the porphyritic granodiorite body (unit 5b).

c) An approximately north-trending fault which cuts the folded megacrystic granite body (unit 6) in the northwestern part of the area.

There is no geological evidence for the other faults indicated on the maps of Padgham (1968) and Macdonald and Posehn (1976). Their existence is not required to explain distribution of rocks and lithological boundaries. If present at all, they have insignificant displacement.

3. Metamorphism

Granoblastic fabrics typify most rocks in the area, suggesting major recrystallization of the main rock-forming minerals after the last significant deformation event (D_2). Indicator minerals andalusite, sillimanite (in the west), muscovite, garnet, hornblende and diopside signify low- tc mid-amphibolite facies metamorphic conditions, with highest grades being attained in the west. Most of these minerals appear to have been in equilibrium during the post-D_2 recrystallization event, but textural evidence suggests that some phases formed earlier in the deformation history. For example, sillimanite-quartz faserkiesel (possibly with retrograde muscovite) and elongate garnet porphyroblasts are oriented within the S_1 foliation; their growth was clearly influenced by this event. Andalusite porphyroblasts obviously overgrow S_1 but are locally rotated and internally folded by F_2 crenulation, suggesting formation between the D_1 and D_2 events. Post-S_1 hornblende porphyroblasts in mafic gneisses are locally augened by reactivation along S_1, again suggesting pre-D_2 growth. Change from monomineralic quartz veining to quartz-K feldspar veins in metasediments in the northwest may indicate incipient partial melting in this area prior to or during D_1.

4. Economic Geology

The Northern Lights volcanics have been described by Macdonald and Posehn (1976) as lithologically similar to volcanics at Hanson Lake, which have most recently been mapped by Sibbald (1989) and Koziol and Ostapovitch (1990). In the Hanson Lake area, base metal mineralization at the Western Nuclear Mine and McIlvenna Bay deposit is clearly linked to rhyolite doming occurring at the end of a period dominated by dacitic volcanism. There is widespread evidence of hydrothermal alteration in the mineralized packages and of chemical sedimentation of mixed facies iron formations and calc-silicates in laterally equivalent and overlying rocks (Maxeiner and Watters, this volume).

The Northern Lights volcanics comprise pillowed mafic flows and tuffs and mixed volcaniclastics, which are on average andesitic to dacitic in composition. Rhyolite units are uncommon and restricted to thin intervals in the mafic flows and mixed volcaniclastics. They probably represent crystal tuffs. The mixed volcaniclastics are interpreted here as subaqueously deposited distal volcanic and/or reworked volcanic material. They lack evidence of alteration and do not contain chemical sediments or fine argillaceous intervals marking periods of slow sedimentation. This is in marked contrast to the host rocks of the Bigstone Lake deposit some 10 km to the south (Adamson, 1991). Subvolcanic intrusions which characterize Hanson Lake are also noticeably absent in the Northern Lights segment. Based on these parameters the mapped area appears to be relatively unattractive for base metal exploration, which to date has identified only minor low grade copper-zinc mineralization in the north Limestone Lake-Hand Lake area and north and south of the old Hanson Lake highway between miles 132 and 134 (Padgham, 1968). In the former location, showings are drill indicated, in the latter are concordant gossaned iron sulphides and minor chalcopyrite. Molybdenite and galena are also noted in assessment reports.

5. Discussion and Regional Implications

Several conclusions of regional significance can be drawn from results of current work in the Southeast Arm-Tulabi Brook area.

1. Ductile high-strain, locally mylonitic fabrics throughout the area represent an early stage of structural evolution that are deformed about regional D_2 northerly folds. Thus, they cannot be directly related to movement along the Tabbernor zone; rather, they reflect a pre-existing high-strain zone (or anastomosing zones), later deflected into the Tabbernor zone.

2. The indefinitely delimited belt of highly-strained rocks extending through Southeast Arm and along the eastern limb of the Unser Lake fold may mark an extension of the Hartley Lake shear zone defined farther west in southern Glennie Domain (Delaney, 1988). Highly strained rocks along the southeastern limb of the Sarginson Lake fold possibly represent a subsidiary high-strain zone.

   In the Twin Bays area, Deschambault Lake, there is well-defined northward strain increase in granodioritic rocks forming part of the Oskikebuk complex, though the northern margin of this plutonic body is unexposed. Quartzofeldspathic gneisses (Unit 3 of Padgham, 1968) to the north of the Oskikebuk complex are very fine grained and intensely strained, but appear to be derived from arkosic metasedimentary protoliths rather than from granitoid plutonic rocks. Immediately to the north of these, very fine-grained, evidently highly-strained laminated amphibolites, similar to those in the Southeast Arm area, are of indeterminate protolith. Features seen across this transect are similar to those documented across the Hartley Lake shear zone farther west, and are here interpreted as representing an eastward extension of that zone. Strike continuity to the east and south suggests that the high strain zone continues into the Southeast Arm, through the package of heterogeneously highly strained sediments, mafic and plutonic gneisses documented here. Regional strike continuity to the northeast of the present area suggests that the high-strain zone may root in the structurally underlying Pelican Slide (Lewry *et al.*, 1989) but spatial and temporal relations between the two remain unclear.

3. Most late faults delineated in previous mapping (Padgham, 1968; Macdonald and Posehn, 1976) either do not exist or have negligible displacement. Aside from the Tulabi Brook fault, none of the faults identified during the mapping this summer show more than 1 to 2 km strike displacement and there is no significant change in lithologies, metamorphic grade or structural state across them. The Tulabi Brook fault is clearly a more substantial feature, with sufficient displacement to produce significant general decrease in metamorphic grade to the west (Macdonald and Posehn, 1976), Nonetheless, some major lithological units seem to occur on both sides and it does not seem to be a fundamental early discontinuity.

   Thus, the Tabbernor zone is essentially a late feature across which there is general continuity of earlier ductile structural elements. Ductile high-strain features broadly coincident with the southern part of the zone are considered to mark much earlier, unrelated refolded shear zones fortuitously deflected into the Tabbernor trend. Sharp metamorphic gradients or discontinuities documented elsewhere across the Tabbernor zone (e.g. Wilcox, 1990) may be largely a function of pre-Tabbernor thrust imbrication and refolding.

4. Late leucogranodiorites, leucogranites and related pegmatites (unit 11) are essentially identical to rocks of the 'Jan Lake' suite in the Tulabi-Jan Lakes region and are probably cogenetic/coeval with the latter. However, they commonly show a weak linear and/or planar fabric which is probably of D_2 (northerly folding) age. Their emplacement may thus have spanned and outlasted late stages of D_2 folding.

6. References

Adamson, D. (1991): Geology of the Bigstone Lake Cu-Zn deposit, northern Saskatchewan; Geological Society of CIM, First Annual Field Conference, Saskatoon, Prog. Abstr., p25.

Delaney, G.D., (1988): Bedrock geological mapping, Brownell Lake area (part of NTS 63M-4 and 63L-13); *in* Summary of Investigations 1988, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 88-4, p8-19.

Koziol, M. and Ostapovitch, G. (1990): The Mcllvenna Bay massive sulphide deposit-a case history; *in* Beck, L.S. and Harper, C.T. (eds.), Modern Exploration Techniques Saskatchewan Geological Society, Spec. Publ. 10, p54-70.

Lewry, J.F., (1990): Bedrock geology, Tulabi-Church lakes area: derivation and significance of porphyroclastic gneisses in the Pelican Window; *in* Summary of Investigations 1990, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 90-4, p36-43.

Lewry, J.F. and Macdonald, R. (1988): Observations on deformation in the Glennie Domain and Hanson Lake Block; *in* Summary of Investigations 1988, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 88-4, p35-41.

Lewry, J.F., Macdonald, R. and Stauffer, M.R. (1989): The development of highly strained rocks in the Pelican Window during high grade metamorphism and pervasive anatexis; *in* Summary of Investigations 1989, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 89-4, p58-65.

Macdonald, R. and Posehn, G.A. (1976): Re-investigation of the Pickerel River and Limestone-Tulabi Lakes area; *in* Summary Report of Field Investigations by the Saskatchewan Geological Survey 1976, Sask. Dep. Miner. Resour., p44-52.

Padgham, W.A. (1968): The Geology of the Deschambault Lake District; Sask. Dep. Miner. Resour., Rep. 114, 92p.

Sibbald, T.I.I. (1989): Base metal deposits and geology of the Early Proterozoic Hanson Lake metavolcanics; *in* Summary of Investigations 1989, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 89-4, p66-70.

Wilcox, K.H. (1990): Bedrock geology and gold metallogenesis, Nielson Lake area, Tabbernor fault zone; *in* Summary of Investigations 1990, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 90-4, p74-83.

Glennie-Kisseynew P-T Study 1991: Fieldwork and Preliminary Results[1]

Dexter Perkins[2]

Perkins, D. (1991.: Glennie-Kisseynew P-T study 1991: Fieldwork and preliminary results; *in* Summary of Investigations 1991, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 91-4.

Detailed investigation of metamorphism in the southern part of the Kisseynew Domain, northeastern Saskatchewan commenced in August 1987 (Figure 1). Further fieldwork was conducted during the summers of 1988, 1989 and 1991. Thin sections were made of selected samples collected during 1987-89, and mineral chemical analyses were obtained using a scanning electron microscope/electron microprobe.

Application of conventional thermobarometers has permitted semi-quantitative evaluation of metamorphic pressures and temperatures (Perkins, in press). Temperatures of about 575°C are recorded by biotite-garnet pairs (Perchuk and Lavrent'eva, 1983) near the boundary between the Kisseynew Domain and the Flin Flon Domain. Temperatures increase to the north, reaching around 700°C in the vicinity of Sandy Bay (Figure 2).

Metamorphic pressures, calculated from assemblages of garnet-plagioclase-sillimanite/kyanite-quartz (Newton and Haselton, 1981), are between 4 and 6 kb over most of the area studied. Although some ambiguity remains, it appears that metamorphic pressures in the southern part of the study area are slightly greater than those in the northern part. Several samples from around the borders of the Attitti Block tectonic sheet yield pressures of 8 to 12 kb.

The anomalous pressures suggest that the Attitti Block tectonic sheet may represent exposure of a much deeper structural level than the surrounding Kisseynew Gneisses. Further studies were commenced in August 1991, in order to clarify this relationship, and to obtain more detailed information regarding the pressure and temperature variations near the Kisseynew-Flin Flon domainal boundary.

1. Region of Interest

The area covered by 1991 fieldwork is approximately rectangular, comprising all the area covered by NTS area 63M-1, the northern third of 63L-16, and small portions of 63M-2, 63L-15, 63N-4, 63K-13. In this area, previous geological mapping has been done by Byers and Dahlstrom (1954), Cheesman (1956), Pyke (1961) and Ashton (1986), Ashton *et al.* (1987), and Wilcox (1989). Many samples were collected from the shores of lakes and rivers, including Attitti Lake, Kakinagimak Lake, Wildnest Lake, Mari Lake, Johnson Lake, Annabel Lake and Weetago Bay. The majority were taken from units identified as Missi arenites or Amisk wackes (Ashton *et al.*, 1986). A few of the more mafic units were also sampled. Of over 400 samples collected, thin sections have been made for about 45, and more are being prepared.

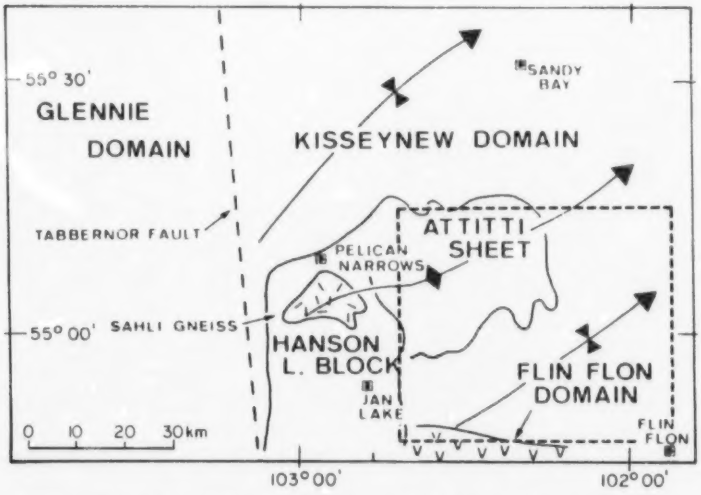


Figure 1 - Sketch map showing area of investigations from 1987-90, and area reinvestigated during the 1991 summer (dashed lines).

2. Lowest Grade Rocks

The lowest grade rocks were found in the vicinities of Annabel and Johnson Lakes, south of the Hanson Lake Road (Highway 106). Many of the metasediments on the shores of both lakes contain muscovite. These lakes are just a few kilometres north of the isograd marking the first appearance of the sillimanite-almandine-biotite assemblage,

---

[1] Saskatchewan Project A.124, from which this report derives, was funded in 1991 under the Canada-Saskatchewan Partnership Agreement on Mineral Development 1990-95.

[2] Department of Geology and Geological Engineering, University of North Dakota, Grand Forks, North Dakota, 58202, USA.

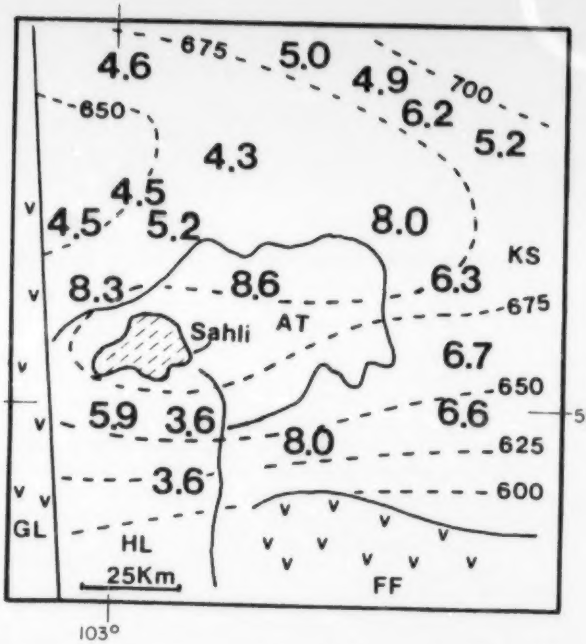


Figure 2 - Smoothed results of thermobarometry (modified from Perkins, in press). Contours give temperature (°C), large numbers are pressure. For clarity, not all calculated pressures are shown. AT: Attitti sheet, Sahli: Sahli charnockite, FF: Flin Flon Belt, GL: Glennie Domain.

which is conveniently used to delineate the boundary between the Flin Flon and Kisseynew Domains (Ashton, pers. comm.). Sillimanite was the only Al_2SiO_5 mineral identified at Annabel and Johnson Lakes. Previous studies report andalusite 5 km farther south (in the Flin Flon Domain), and kyanite has been reported from one sample at Welsh Lake, 6 km west-southwest of Annabel Lake (Ashton *et al.*, 1987; Wilcox, 1989; Ashton, pers. comm.).

3. Highest Grade Rocks

With the exception of the low grade rocks described above, most samples collected in 1991 are upper-amphibolite facies gneisses. Typical major minerals in metasediments are quartz-plagioclase-biotite-garnet-sillimanite. More mafic rock types contain hornblende-garnet-plagioclase-quartz. The highest metamorphic grades appear to have been reached in the Galbraith-McCall-Grindley Lakes area, between Kakinagimak and Attitti Lakes. The metasediments in that area are highly migmatized and disrupted by large volumes of diatexite. The Al_2SiO_5 minerals were not identifiable in hand samples of these highly deformed rocks.

4. Kyanite-bearing Samples

Ashton and Wheatley (1986) and Ashton *et al.* (1986) reported kyanite in 8 to 10 locations in the southern Kisseynew Domain. Perkins (in press) found kyanite at two additional locations. It is important to know the nature and extent of kyanite occurrences, because such information places valuable constraints on metamorphic pressure. Detailed examination was made of most locations where kyanite has been previously reported, and attempts were made to trace kyanite-bearing units on the basis of Ashton *et al.*'s (1986) map. Unfortunately, no new kyanite occurrences were found. Ashton and Wheatley (1986) and Ashton *et al.* (1986), using textural criteria, suggested the kyanite was relict. Perkins (in press) has shown that metamorphic pressure-temperature attained by the kyanite-bearing samples may have been about 6.5 kb, 625°C, that is, just into the kyanite stability field. More study is needed before the kyanite occurrences will be completely understood.

5. Geothermometry

Metamorphic temperatures were previously calculated for 15 samples from the study area and pressures for 5 samples (Perkins, in press). The previous results give a general idea of regional variation in pressure and temperature, but more data are needed to pin down details. Mineral analyses are being obtained for another 27 samples at the present time and more may be obtained this winter. It appears that metamorphic temperatures generally increase from 575°C to 675°C from south to north across the study area. East-west variation in temperature and regional variations in pressure are not presently tightly constrained.

6. Ongoing Investigations

During the winter of 1991-92, new results should place much better constraints on regional variation in metamorphic grade. The results will be compared with previous and ongoing mapping of Ashton and Leclair, this volume) in order to determine the relationship between metamorphic grade and regional structures. In order to fill in gaps, more areas will most likely have to be visited in 1992. In particular, the areas between Little Mari and Wildnest Lakes, and between Kakinagimak and Mari Lakes, contain Amisk wacke and Missi arenite that have yet to be sampled in detail. Farther west, north of Granite Lake, there are several small lakes that should also be sampled.

7. References

Ashton, K.E. and Wheatley, K.J. (1986): Preliminary report on the Kisseynew gneisses in the Kisseynew-Wildnest Lakes area, Saskatchewan; *in* Current Research, Geol. Surv. Can., Pap. 86-1B, p305-317.

Ashton, K.E., Wheatley, K.J., Moser, D., Paul, D., and Wilcox, K.H. (1986): The Kisseynew gneisses of Saskatchewan: update; *in* Summary of Investigations 1986, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 86-4, p145-158.

Ashton, K.E., Wilcox, K.H., Wheatley, K.J., Paul, D., and de Tombe, J. (1987): The boundary zone between the Flin Flon Domain, Kisseynew Gneisses and Hanson Lake Block, northern Saskatchewan; *in* Summary of Investigations 1987, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 87-4, p131-134.

Byers, A.R. and Dahlstrom, C.D.A (1954): Geology and mineral deposits of the Amisk-Wildnest Lakes area, 63L-9, 63L-16 Saskatchewan; Sask. Dept. Miner. Resour., Rep. 14, 177p.

Cheesman, R.L. (1956): The Geology of the Mari Lake area Saskatchewan; Sask. Dep. Miner. Resour., Rep. 23, 40p.

Newton, R.C. and Haselton, H.T. (1981): Thermodynamics of the garnet-plagioclase-Al_2SiO_5-quartz geobarometer; *in* Saxena, K.K. (ed.), Thermodynamics of Minerals and Melts, Advances in Physical Geochemistry, v1, p131-147.

Perkins, D. (in press): Metamorphism of the Kisseynew gneisses and related rocks of the Reindeer Zone, Trans-Hudson Orogen, northern Saskatchewan; Can. J. Earth Sci.

Perchuk, K.L. and Lavrent'eva, I.V. (1983): Experimental investigation of exchange equilibria in the system cordierite-garnet-biotite; *in* Saxena, K.K. (ed.), Kinetics and Equilibrium in Mineral Reactions, Advances in Physical Geochemistry, v1, p199-240.

Pyke, M.W. (1961): The geology of the Attitti Lake area (west half), Saskatchewan; Sask. Dep. Miner. Resour., Rep. 54, 33p.

Wilcox, K.H. (1989): Investigations of Missi metasedimentary rocks in the Amisk-Welsh Lakes area, Saskatchewan; *in* Current Research, Geol. Surv. Can. Pap. 89-1C, p165-171.

Metamorphic P-T Studies in the Jan Lake Area, Hanson Lake Block[1]

Syed Abbas-Hasanie[2], John F. Lewry[2] and Dexter Perkins[3]

Abbas-Hasanie, S., Lewry, J.F. and Perkins, D. (1991): Metamorphic P-T studies in the Jan Lake area, Hanson Lake Block; *in* Summary of Investigations 1991, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 91-4.

Recent work (Lewry and Macdonald, 1988; Lewry *et al.* 1989,1990; Lewry, 1990) suggests that highly strained gneisses comprising the 'Pelican Slide', overlying and enveloping the Sahli monzocharnockite in the Hanson Lake Block, represent a fundamental, high-grade metamorphic, ductile mylonitic 'sole zone' separating Archean basement and overlying Early Proterozoic allochthons. Better documentation of the kinematic, thermobaro-metric and isotopic history of rocks in this high-strain zone is critical to an overall understanding of tectonic evolution in the Reindeer Zone of the Trans-Hudson Orogen. The objective of this study is to better establish the metamorphic history of rocks in the Pelican Slide. Part of that history can be established by geothermometric and geobarometric studies of critical metamorphic mineral parageneses in suitable rocks. Results will be correlated with parallel geochemical and isotopic studies being conducted at the University of Saskatchewan (Kerrich, this volume) and detailed fieldwork by Shi Rupan (Shi and Lewry, this volume).

During the first two weeks of August 1991, the first author collected over 100 samples of both paragneisses and orthogneisses from lakeshore outcrops around Jan Lake and southwest Mirond Lake, with special emphasis on sampling pelitic metasediments. Appropriate calibrations of standard cation exchange geothermometers and geobarometers will be applied to selected samples of pelitic rocks containing garnet, biotite, sillimanite, plagioclase, quartz, cordierite and K-feldspar. Chemical analysis of minerals will be carried out by electron microprobe at the University of North Dakota. Metamorphic temperatures and pressures derived from application of the garnet-biotite and garnet-cordierite thermometers of Perchuk and Lavrente'eva (1983) and grossular-anorthite-sillimanite-quartz barometer of Newton and Haselton (1981) will be presented after completion of microprobe mineral analysis.

This report gives details of sample collection and preliminary observations on metamorphism in the Jan lake area, based on fieldwork and initial thin-section examination.

1. Regional Background

Recent models (Lewry *et al.*, 1990: Bickford *et al.*, 1990) suggest that the mostly juvenile Reindeer Zone underwent early thrust imbrication and nappe development with coeval development of localized, gently inclined, ductile high-strain zones, followed by polyphase refolding. In the Jan-Mirond Lakes area, the early event appears to be represented by a compound progressive strain, S_1/S_2 layer-parallel foliation and locally preserved intrafolial isoclinal folds probably of more than one generation (cf. Lewry *et al.* 1989). Over much of the region, this early planar fabric is essentially mylonitic in many rock types. Unequivocal evidence of the D_1 event is only locally preserved. D_2 deformation included extensive development of generally tight to isoclinal mesoscopic to microscopic folds in most rock types and a pervasive, regionally developed S_2 axial plane schistosity which generally overprints and obliterates S_1 except in hinge zones of some D_2 folds.

This early progressive strain sequence evidently occurred under high- to very high-grade metamorphic conditions, as demonstrated by extensive anatexis of pelitic to psammopelitic paragneisses, broadly coeval development of fibrolitic to prismatic sillimanite-K feldspar-garnet-bearing assemblages, and absence of primary muscovite in pelitic to psammopelitic rocks. Migmatization of both metasedimentary and meta-igneous rocks of felsic to intermediate compositions during both D_1 and D_2 is evident throughout the area (cf. Lewry *et al.*, 1989).

D_1/D_2 structures are refolded about broad open to tight semi-upright folds during D_3 and D_4 to produce the dominant major structures in the region, including the complex, elongate, F_3/F_4 interference dome, cored by the Archean Macmillan Point and Sahli basement 'windows' (Lewry, *et al.*, 1990). F_3 folds in the study area are north- to northwest-trending open to tight folds, locally refolded by generally upright, variably plunging, open F_4 folds about northeast-trending axial surfaces. The interpretative geological section through the Hanson Lake Block and adjacent domains presented by Lewry *et al.* (1990) indicates that basement rocks may underlie much of the adjacent Kyaska nappe sheet (Kisseynew Domain) and Flin Flon Domain and are exposed in the core of the Hanson Lake Block solely as the result of F_3/F_4 refolding and uplift.

---

(1) Saskatchewan Project A.125, from which this report derives, was funded in 1991 under the Canada-Saskatchewan Partnership Agreement on Mineral Development 1990-95.

(2) Department of Geology, University of Regina, Regina, Saskatchewan.

(3) Department of Geology and Geological Engineering, University of North Dakota, Grand Forks, North Dakota.

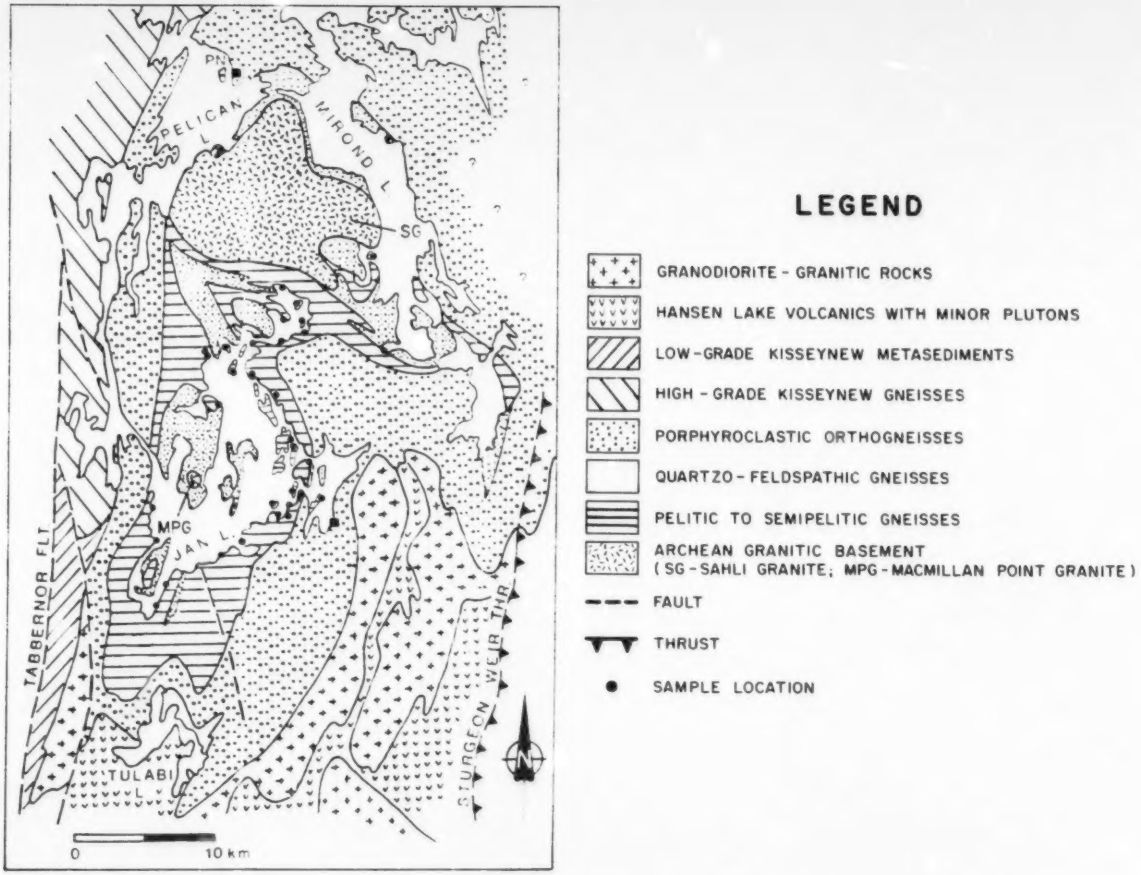

PN
PELICAN L.
MIROND L.
SG
MPG
JAN L.
TABBERNOR FLT
STURGEON WEIR THR.
TULABI L.
N
0
10 km
LEGEND
GRANODIORITE - GRANITIC ROCKS
HANSEN LAKE VOLCANICS WITH MINOR PLUTONS
LOW-GRADE KISSEYNEW METASEDIMENTS
HIGH-GRADE KISSEYNEW GNEISSES
PORPHYROCLASTIC ORTHOGNEISSES
QUARTZO-FELDSPATHIC GNEISSES
PELITIC TO SEMIPELITIC GNEISSES
ARCHEAN GRANITIC BASEMENT
(SG-SAHLI GRANITE; MPG-MACMILLAN POINT GRANITE)
FAULT
THRUST
SAMPLE LOCATION

Figure 1 - General geology of the Hanson Lake Block and environs (after Lewry et al., 1989) showing general sample locations for this study. PN=Pelican Narrows.

2. Metamorphic History

Locations of samples collected this summer are shown in Figure 1. To date, only a few of these samples have been examined in thin section. Mineral assemblages so far documented in pelitic to psammopelitic rocks include:

Quartz-plagioclase-biotite-garnet-sillimanite-K feldspar-cordierite

Quartz-plagioclase-biotite-garnet-sillimanite-K feldspar

Quartz-plagioclase-biotite-garnet-sillimanite

Quartz-plagioclase-biotite-garnet-cordierite

Quartz-plagioclase-biotite-garnet

Primary muscovite was not seen. Such assemblages, in conjunction with extensive partial melting, indicate general upper amphibolite facies metamorphism. However, a possible earlier attainment of higher pressure granulite facies metamorphism is implied by previously documented preservation of relict spinel in cores of some cordierite crystals in pelites and the occurrence of orthopyroxene in a metabasite sill east of MacMillan Point (Craig, 1989). Trace amounts of green spinel have been observed in only one thin section related to the present study. Preservation of a spinel-sillimanite-garnet-quartz assemblage in silica saturated rocks suggests the attainment of relatively high pressure granulite facies metamorphic conditions (e.g. Annerson and Siefert, 1981). Sparse documentation of kyanite in high-grade pelitic gneisses in the Jan Lake area (Perkins, in preparation) may also reflect such an early high pressure event.

Observed textural relations indicate two phases of cordierite growth in the pelite assemblages. In the first, cordierite appears to have developed coevally with garnet via the prograde reaction:

sillimanite + biotite + quartz = K-feldspar + garnet + cordierite + H_2O

P-T conditions of metamorphism for this reaction are constrained by the experimental work of Holdaway and Lee (1977). Depending on fluid pressure, possible temperatures and confining pressures for the reaction range from 650° to 800°C and 3 to 7 Kb. Since these pelitic rocks underwent anatexis, P_{H_2O} was probably less than P_{Total} during prograde metamorphism. Moreover, the presence of graphite, noted in some pelitic assemblages, implies dilution of the fluid phase by addition of CO_2.

Elsewhere, textural relations suggest that later cordierite has partially replaced garnet and sillimanite. In this case, it may have formed by the following decompression reaction (Barber and Yardley, 1985):

garnet + sillimanite + quartz + H_2O = cordierite

It is clear that peak metamorphic conditions in the Hanson Lake Block may have, at least locally, attained high pressure granulite facies, with subsequent decline to lower pressure upper amphibolite facies conditions. The early high pressure, high temperature metamorphism indicated by the presence of kyanite and spinel in pelitic rocks may be related to crustal thickening due to nappe emplacement (Lewry, *et al.* 1990). It should be possible to relate changes in P-T conditions of metamorphism to deformational events by textural evidence. At present, it is tentatively suggested that Early Proterozoic high-pressure granulite facies metamorphism, including that of the Sahli basement (Craig, 1989), was initiated during nappe development in the Reindeer Zone and persisted throughout the D_2 event. It has been suggested that high-grade regional metamorphism in the Reindeer Zone outlasted the D_3 deformation (Lewry, *et al.* 1990), a view supported by field observations made during the present study. It is possible that decompression began during D_3 regional folding, contemporaneous uplift and erosion.

3. References.

Annerson, H. and Siefert, F. (1981): Stability of the assemblage orthopyroxene-sillimanite-quartz in the system $MgO\text{-}FeO\text{-}Fe_2O_3\text{-}Al_2O_3\text{-}SiO_2\text{-}H_2O$; Contrib. Mineral. Petrol., v77, p158-165.

Barber, J.P. and Yardley, W.D. (1985): Conditions of high-grade metamorphism in the Dalradian of Connemara, Ireland; J. Geol. Soc., v142, Pt1, p87-96.

Bickford, M.E., Collerson, K.D., Lewry, J.F., Van Schmus, W.R. and Chiarenzelli, J.R. (1990): Proterozoic collisional tectonism in the Trans-Hudson Orogen, Saskatchewan; Geol., v18, p14-18.

Craig, L.G. (1989): Geology of the Pelican Narrows area of east-central Saskatchewan; unpubl. Ph.D. thesis, Univ., of Sask., 254p.

Holdaway, M.J. and Lee, S.M. (1977): Fe-Mg cordierite stability in high-grade pelitic rocks based on experimental, theoretical and natural observations; Contrib. Mineral. Petrol., v63, p175-198.

Lewry, J.F. (1990): Bedrock Geology, Tulabi-Church Lakes area: derivation and significance of porphyroclastic gneisses in the Pelican Window; *in* Summary of Investigations 1990, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 90-4, p36-43.

Lewry, J.F. and Macdonald, R. (1988): Observations on deformation in the Glennie Domain and Hanson Lake Block; *in* Summary of Investigations 1988, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 88-4, p35-41.

Lewry, J.F., Macdonald, R. and Stauffer, M.R. (1989): The development of highly strained rocks in the Pelican Window during high-grade metamorphism and pervasive anatexis; *in* Summary of Investigations 1989, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 89-4, p58-65.

Lewry, J.F., Thomas, D.J., Macdonald, R. and Chiarenzelli, J. (1990): Structural relations in accreted terranes of the Trans-Hudson orogen, Saskatchewan: telescoping in a collisional regime?; *in* Lewry, J.F. and Stauffer, M.R.(eds.), The Early Proterozoic Trans-Hudson Orogen of North America; Geol. Assoc. Can., Spec. Pap. 37, p76-94.

Newton, R.C. and Haselton, M.T. (1981): Thermodynamics of garnet-plagioclase-Al_2SiO_5-quartz geobarometer; *in* Newton, R.C. and others, (eds.), Thermodynamics of Minerals and Melts; Springer-Verlag, New York.

Perchuk, L.L. and Lavrent'eva, I.V. (1983): Experimental investigation of exchange equilibria in the system cordierite-garnet-biotite; *in* Saxena, S.K., (ed.), Kinetics and Equilibrium in Mineral Reactions, Springer-Verlag, New York, p199-239.

Lithogeochemical Studies, Wildnest-Tabbernor Area[1]

G.R. Parslow[2]

Parslow, G.R. (1991): Lithogeochemical studies, Wildnest-Tabbernor area; *in* Summary of Investigations 1991, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 91-4.

Over the period 1979-86 detailed geochemical studies were carried out by the author and a coauthor in the East Amisk, Annabel, Hamell, Hanson, Mari and Wildnest Lakes areas (Parslow, 1986; Parslow and Gaskarth, 1984, 1985, 1986, 1987; Parslow *et al.*, 1986). Results for the greenschist facies rocks of the first three areas allowed for a clear identification of the environment of formation of the volcanic rocks: predominantly tholeiitic arc volcanics with evidence of some MORB component. Data for the upper amphibolite facies rocks of the last two areas gave more equivocable interpretations, probably due to the lack of good geological control of the sampling, although the effects of metamorphism cannot be ruled out.

The present study has three major objectives:

a) to reinterpret the Kisseynew Domain analyses in light of recent remapping of the southern portion of the Domain (Ashton and Wheatley, 1986; Ashton *et al.*, 1987);

b) to attempt to characterize the lithogeochemical features of mafic and felsic volcanic rocks in the Wildnest-Tabbernor Transect area in conjunction and in consultation with Ashton who is remapping the area (Ashton, this volume);

c) to attempt to find geochemical indicators over mineralized zones within the Wildnest-Tabbernor Transect area, as a guide for further exploration work in the region.

To date, samples from the earlier geochemical studies in the Kisseynew Domain have been reclassified according to Ashton and Wheatley's (1986) and Ashton *et al.*'s (1987) lithological and stratigraphic subdivisions. Of some 290 samples collected about 260 were categorized using sample locations and recent geological maps. The remaining samples will be classified in the near future using archival material stored at Saskatchewan Energy and Mines' Subsurface Laboratory. An apparent consistent error of approximately one kilometre in the UTM east values, digitized in-house by SEM, will be addressed and corrected in the coming months.

Preliminary studies of the major and trace element data suggest that a geochemical distinction between Amisk wackes and Amisk mafic volcanics is possible. On the other hand, distinction between Amisk mafic volcanics and later post-Missi mafic intrusives is not clear and requires more study.

This summer about 75 samples were collected in the Wildnest-Attitti Lakes area of the Wildnest-Tabbernor Transect (Ashton and Leclair, this volume). The samples include a variety of mafic to felsic gneisses tentatively interpreted as high grade volcanic rocks and minor subvolcanic intrusions. In addition, some 10 samples of altered volcanics were sampled in order to carry out mass balance calculations vis-a-vis fresh volcanics. At the present time these samples are being prepared for major and trace element analysis. Once these analyses have been completed, a small number of the freshest samples will be subjected to rare earth element (REE) analysis in order to characterize the volcanic rocks. This study will include an attempt to determine the origin of samples tentatively interpreted as felsic volcanic rocks. Samples from outcrops with visible sulphide minerals will be analyzed for base metals and Au, in addition to the above-mentioned major, trace, and REE analyses.

1. References

Ashton, K.E. and Wheatley, K.J. (1986): Preliminary report on the Kisseynew Gneisses in the Kisseynew-Wildnest Lakes area, Saskatchewan; *in* Current Research, Pt. B, Geol. Surv. Can., Pap. 86-1B, p305-317.

Ashton, K.E., Wilcox, K.H., Wheatley, K.J., Paul, D. and de Tombe, J. (1987): The boundary zone between the Flin Flon domain, Kisseynew gneisses and Hanson Lake block, northern Saskatchewan; *in* Summary of Investigations 1987, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 87-4, p131-134.

Parslow, G.R. (1986): Kisseynew metallotect geochemical study; *in* Summary of Investigations 1986, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 86-4, p84-96.

Parslow, G.R. and Gaskarth, J.W. (1984): Geochemistry of the east Amisk area; Sask. Energy Mines, Open File Rep. 84-23, 156p.

__________ (1985): Kisseynew metallotect geochemical study; *in* Summary of Investigation 1985, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 85-4, p50-58.

---

(1) Saskatchewan Project A.126, from which this report derives, was funded in 1991 under the Canada-Saskatchewan Partnership Agreement on Mineral Development 1990-95.

(2) Department of Geology, University of Regina, Regina, Saskatchewan.

__________ (1986): Geochemistry of the Hanson Lake area; Sask. Energy Mines, Open File Rep. 86-1, 122p.

__________ (1987): Geochemistry of the Harrell Lake area; Sask. Energy Mines, Open File Rep. 87-4, 73p.

Parslow, G.R., Gaskarth, J.W. and Zhang, J.L. (1986): Geochemistry of the Annabel Lake area; Sask. Energy Mines, Open File Rep. 86-3, 154p.

U-Pb Geochronological Investigations in the Trans-Hudson Orogen, Saskatchewan: Preliminary Results by the ROM Laboratory in 1990-91[1]

L.M. Heaman[2], S.L. Kamo[2], G.D. Delaney, C.T. Harper, B.A. Reilly, W.L. Slimmon, and D.J. Thomas

---

Heaman, L.M., Kamo, S.L., Delaney, G.D., Harper, C.T., Reilly, B.A., Slimmon, W.L. and Thomas, D.J. (1991): U-Pb geochronological investigations in the Trans-Hudson Orogen, Saskatchewan: Preliminary results by the ROM Laboratory in 1990-91; *in* Summary of Investigations 1991, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 91-4.

This paper summarizes the U-Pb geochronological results for collaborative projects initiated in 1990-91. In addition to the 28 samples collected in August of 1990, an additional set of 17 samples was collected in August of 1991 to expand sampling coverage, especially in the La Ronge and Flin Flon Domains. A summary of the preliminary U-Pb age results obtained in 1991 is presented in Table 1 with uncertainties quoted at 2 percent. Details of the analytical procedures can be found elsewhere (Krogh, 1973; 1982; Heaman and Machado, in press). The following discussion is focussed on areas where extensive sampling initiatives were undertaken. The results not discussed here will be presented in forthcoming papers.

1. Glennie Domain

a) Gee Lake-Sadler Lake Area

Four samples were selected from this area for a U-Pb study to provide constraints on the timing of volcanism, sedimentation, metamorphism and deformation. The U-Pb zircon ages for two rhyolite flows (8930-GC1 and 9030-1) from different structural packages are 1866 ±3 Ma and 1875 ±3 Ma, respectively. A U-Pb age of 1795 ±1 Ma for metamorphic monazite isolated from a pelitic schist (9030-2) at Grand Rapids is interpreted as the best estimate for the timing of metamorphism in this area. The youngest age for a detrital zircon from this same sample is 1837 Ma and represents a maximum age for the timing of sedimentation. An additional zircon analysis from the post-tectonic aplite sample 8930-GC2 (results for three zircon fractions presented in Delaney *et al.*, 1990) has the youngest $^{207}Pb/^{206}Pb$ age of 1856 Ma. Although there is evidence for zircon inheritance in this sample, the results for these four fractions taken together indicate that the aplite could be as young as 1760 Ma.

2. Flin Flon Domain

a) Mystic Lake Area

Two samples, a foliated granodiorite (9023-0744) and a rhyolite porphyry dike (9023-0743), collected in the Mystic Lake area were investigated to constrain the timing of felsic magmatism. Preliminary results for two zircon analyses from the granodiorite sample, indicate an age

Table 1 - Summary of Preliminary U-Pb Results

Sample No.	*Rock Type*	*Location*	*Domain*	*UTM E*	*UTM N*	*n*	*UI*[1]	*LI*	*Collab.*
8812-2000	Diabase	Bartlett Lake	La Ronge	500550	6153540	?	z	1885 ±3	Thomas
9011-201	Diorite	Tower Lake	La Ronge	559330	6222760	4	z	1874 ±1	Harper
9022-92	Rhyolite	Porky Lake	Glennie	586620	6173820	4	z	1838 ±2	Delaney
9022-94	Porphyry	Maynard Lake	Glennie	571460	6096540	4	z	1834 ±4	Delaney
9030-1	Rhyolite	Gee Lake	Glennie	608550	6115880	1	z	1875	Slimmon
9030-2	Semipelite	Grand Rapids	Glennie	571520	6134600	2	m	1795 ±1	Slimmon
8930-GC1	Rhyolite	Gee Lake	Glennie	583710	6123470	3	z	1866 ±3	Slimmon
8930-GC2	Aplite	Gee Lake	Glennie	580390	6124810	4	z	2112-1760?	Slimmon
9023-0743	Porphyry Dike	Mystic Lake	Flin Flon	308910	6059950	2	z	1797min	Reilly
9023-0744	Granodiorite	Mystic Lake	Flin Flon	309165	6060575	2	z	1908 ±3	Reilly
9012-0369	Gabbro	Phantom Lake	Flin Flon	315770	6065500	4	z	1834 ±4	Thomas
9012-371	Granodiorite	Phantom Lake	Flin Flon	315770	6065500	4	z	1841 ±3	Thomas

Note: [1] Letter designation corresponds to mineral analysed: z = zircon, m = monazite, min = minimum age; n refers to the number of fractions analysed

---

(1) Saskatchewan project A.131, from which this report derives, was funded in 1991 under the Canada-Saskatchewan Partnership Agreement on Mineral Development 1990-95.

(2) Jack Satterly Geochronology Laboratory, Royal Ontario Museum, 100 Queen's Park, Toronto, Ontario, M5S 2C6

of 1908 ±3 Ma which represents one of the oldest intrusive units presently recognized in the Flin Flon Domain. The U-Pb zircon result for the least discordant fraction (9 percent) from the rhyolite porphyry dike has a $^{207}Pb/^{206}Pb$ age of 1797 Ma and is interpreted as the minimum age for dike emplacement. Additional sample collecting in the Mystic Lake-Wekach Lake areas in August 1991 has been completed with the intent of establishing the extent of the pre-1900 Ma old intrusive rocks in this region.

b) Phantom Lake-Boot Lake Area

Samples of intrusive rocks in the Phantom Lake-Boot Lake area were selected in an attempt to establish the exact timing of the Boundary, Boot Lake and Phantom Lake intrusions. Samples of all three intrusive units were collected from the same outcrop area where field relationships are well documented. At this sampling site the relative age of intrusion, from oldest to youngest, is Boundary Intrusion gabbro (9012-0369), Boot Lake monzodiorite and Phantom Lake granodiorite (9012-0371). The U-Pb zircon/sphene age obtained for the Boundary Intrusion of 1834 ±4 Ma and the U-Pb zircon age obtained for the Phantom Lake granodiorite of 1841 ±3 Ma are not consistent with the observed field relationships. The accuracy of the Boundary Intrusion age is poorly constrained by a slightly discordant sphene analysis and two very discordant (7 and 10 percent) zircon analyses. Other samples of Boundary Intrusion gabbro were collected in 1991 in an attempt to obtain a more reliable and accurate emplacement age for these intrusions.

Additional sampling in the Flin Flon Domain was completed in 1991 to attempt direct U-Pb dating of volcanic units in the vicinity of Flin Flon. Samples of high-grade metavolcanic and metaplutonic rocks from the southern margin of the Flin Flon Domain were also collected in 1991 in collaboration with K. Ashton in order to compare the timing of volcanism and plutonism between the lower and higher grade regions of the Flin Flon Domain.

3. References

Delaney, G.D., Heaman, L.M., Kamo, S., Parrish, R.R., Slimmon, W.L., and Reilly, B.A. (1990): U-Pb sphene/zircon geochronological investigations; *in* Summary of Investigations 1990, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 90-4, p54-57.

Heaman, L.M. and Machado, N. (in press): Timing and nature of midcontinent rift alkaline magmatism, North America: Evidence from the Coldwell Complex; Contrib. Mineral. Petrol..

Krogh, T.E. (1973): A low-contamination method for hydrothermal decomposition of zircon and extraction of U and Pb for isotopic age determinations; Geochim. Cosmochim. Acta, v37, p485-494.

__________ (1982): Improved accuracy of U-Pb zircon ages by the creation of more concordant systems using an air abrasion technique; Geochim. Cosmochim. Acta, v46, p637-649.

The Age of the Mine Granites, Goldfields Area[1]

David S. O'Hanley[2], T. Kurtis Kyser[2] and T.I.I. Sibbald

O'Hanley, D.S., Kyser, T.K. and Sibbald, T.I.I. (1991): The Age of the Mine Granites, Goldfields area; *in* Summary of Investigations 1991, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 91-4.

U-Pb zircon and Rb-Sr whole rock geochronology in the Beaverlodge area of northern Saskatchewan suggests a complex geological history for this region. A compilation of ages from various rock types (Table 1) shows a spread of dates, from 1950 to 3050 Ma, which can be variously divided into several events. For example, at Goldfields in the south central part of the region, at least two episodes of plutonism at ca. 2340 Ma (MacIntosh Bay and Athona Mine granites) and ca. 2000 Ma (Box Mine granite) are implied by U-Pb zircon data (Persons, 1988), whereas field relationships indicate these plutons as contemporaneous (Sibbald, 1984).

The main objective of the present study has been to obtain $^{207}Pb/^{206}Pb$ ages of single zircons using the Pb evaporation technique and thereby resolve the conflict between radiometric and field observations. The new data have: 1) illustrated an inheritance component in the U-Pb zircon ages for the plutons, 2) suggested a single age for the plutons and 3) bracketed the time of deposition of the Murmac Bay Group, which hosts the plutons.

Table 1 - Previous U-Pb Zircon and Whole-rock Rb/Sr Geochronology from the Goldfields Area, Saskatchewan

Sample	Age (Ma)[a]	Method	Reference
Tazin Lake area			
Granitic dyke in older gneiss	2595 ±7	U-Pb	1
Granodiorite xenolith in	2596 ±10	U-Pb	1
granitic dyke	2605 ±10	U-Pb	1
Granitic gneiss, host to dyke	2639 ±16	U-Pb	1
Megacrystic granodiorite	2595 ±4	U-Pb	1
Train Lake Domain			
Quartzofeldspathic gneiss	1950 ±125	Rb/Sr	2
Nevins Lake Block			
Hypersthene gneiss	2630 ±130	Rb/Sr	2
Garnetiferous felsic gneiss	2315 ±60	Rb/Sr	2
Foot Bay gneiss	2513 +36/-22	U-Pb	3
Donaldson Lake gneiss	2179 ±12	U-Pb	3
Lodge Bay granite	2839 ±67	Rb/Sr	4
	3072 ±41	U-Pb	1
Elliott Bay granite	3014 ±10	U-Pb	1
MacIntosh Bay granite	2356 ±14	U-Pb	1
Gunnar mine granite	2181 ±5	U-Pb	1
Box mine granite	1994 ±37	U-Pb	1
Athona mine granite	2327 ±17	U-Pb	1

(a) uncertainty in age is two-sigma value.
Methods are U-Pb: U-Pb zircon; Rb/Sr: Rb/Sr whole-rock.

References:
1. Van Schmus *et al.* (1986);
2. Bell & Blenkinsop (1984);
3. Trembley *et al.* (1981);
4. Bikerman *et al.* (1990).

1. Geological Setting

The Goldfields area is located on the north shore of Lake Athabasca in northern Saskatchewan (Figure 1). Geologically it is situated in the Nevins Lake Block of the Rae Province. The eastern Nevins Lake Block comprises Archean-age granulite facies felsic, garnetiferous and mafic gneisses with rare quartzite influenced by retrogressive amphibolite to greenschist facies metamorphism. In the west, a lower metamorphic grade shelf-type supracrustal assemblage, the Murmac Bay Group, bears an uncertain relationship to the high grade gneisses. It sits unconformably on a ca. 3.0 Ga basement represented by the Lodge Bay and Elliot Bay granites and has been argued both as Archean (Tremblay, 1972) and Early Proterozoic in age (Sibbald and Lewry, 1980).

The Murmac Bay Group comprises amphibolite facies metamorphic equivalents of quartzites, carbonates, shales, iron formation and gabbro (as sills) imprinted by two major deformation events (Sibbald, 1984; Roberts, 1990). These rocks are intruded by a suite of leucogranites called the "North Shore Plutons" which were

---

[1] Saskatchewan Project A.141, from which this report derives, was funded in 1991 under the Canada-Saskatchewan Partnership Agreement on Mineral Development 1990-95 and an operating grant to TKK from NSERC.

[2] Department of Geological Sciences, University of Saskatchewan, Saskatoon, Saskatchewan S7N 0W0

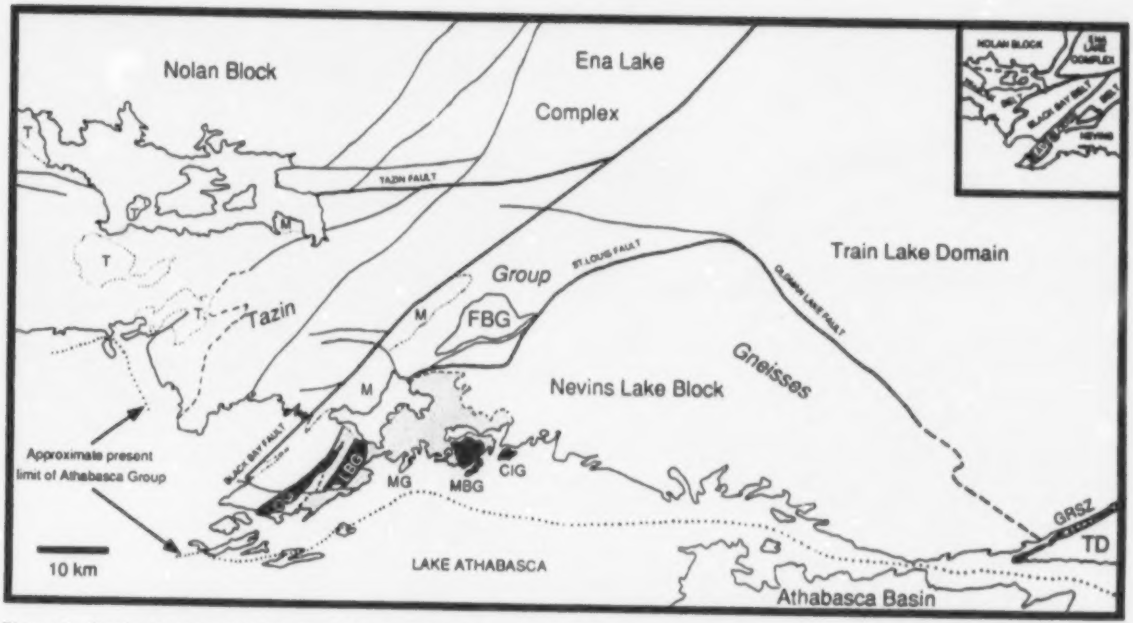


Figure 1 - Geologic map of the Beaverlodge area on the north shore of Lake Athabasca. The Murmac Bay Group is shaded grey. The North Shore plutons are black (GG = Gunnar granite; MG = mine granites; MBG = MacIntosh Bay granite; CIG = Cameron Island granite). LBG = Lodge Bay granite (basement) also black; GRSZ = Grease River Shear Zone; TD = Tantato Domain.

emplaced between the two deformations during peak metamorphism. Among these plutons are the MacIntosh and Gunnar granites and the auriferous, Box, Athona and Frontier "mine granites". Field observations and chemical compositions suggest the plutons grade from products of '*in situ*' granitization (metasomatic replacement granites) to bodies that are intrusive. A suite of 'older' granites characterized by the presence of early first deformation foliations (S_1), marked by preferred orientations of mafic minerals, also occurs in the Goldfields area. All rock types have undergone retrogressive greenschist facies metamorphism.

2. Analytical Techniques and Results

Zircons were separated from samples of the Box, Athona and Frontier mine granites, older granites from near the Box mine granite, and the Frontier quartzite (Quirt and Rees, 1987). The single zircon Pb-evaporation technique as used in this study is identical to that reported by Ansdell and Kyser (1990).

Zircons from the Box and Athona mine granites and one zircon from the Frontier mine granite yield $^{207}Pb/^{208}Pb$ ages within a relatively restricted range from 2170 to 1950 Ma (Table 2 and Figure 2), which is considered to be the age of emplacement. Three other zircons from the Frontier mine granite show ages of 2930, 2510 and 2340 Ma respectively. Detrital zircons from quartzites, adjacent to the Frontier granite, fall within the range from 2380 to 2430 Ma, implying an Early Proterozoic or younger age for these clastic sediments. Based on the ages obtained from the Frontier quartzite the 2510 and 2340 ages from the Frontier granite are considered to be inherited detrital zircons as well. These two zircon ages support the hypothesis that the Frontier granite is metasomatic.

The intervening ages of zircons (2340-2510) in the granite and the quartzites suggest the possibility of a second Late Archean, 2.5 Ga basement source (e.g. Foot Bay gneiss, Nevins Lake Block, Tazin Lake area) (Table 1 and Figure 1). However, the oldest zircon age 2930 is similar to that obtained from basement rocks (Lodge Bay granite) immediately underlying the Murmac Bay Group and it can be attributed to inheritance from a 3.0 Ga basement source.

Six zircons from the two samples of the 'older' granite proximal to the Box mine granite yield variable $^{207}Pb/^{206}Pb$ ages from 2779 Ma to 2964 Ma (Table 2 and Figure 2). Multiple heating steps on several of these zircons produced higher $^{207}Pb/^{206}Pb$ and $^{208}Pb/^{206}Pb$ ratios (and older ages) with increasing temperature implying that older zircons have been variably reset during metamorphism around the outermost portions of the grains. The maximum age of resetting is taken as the age of the youngest zircon, which is 2779 Ma.

In Figure 2 the $^{207}Pb/^{206}Pb$ ages of the zircons are shown divided into three groups. The oldest group, consisting of all of the zircons from the 'older' granites and one zircon from the Frontier mine granite, similar in age to the Archean basement rocks of the Murmac Bay Group, the second group, consisting of zircons from the Frontier quartzite and two zircons from the Frontier mine granite, representing zircons deposited in the Murmac Bay Group, presumably derived from a Late Archean

Table 2 - $^{207}Pb/^{206}Pb$ Ages of Zircons from Lithologies in the Goldfields Area, Saskatchewan. Each age represents an individual zircon.

Sample #	Lithology	Age (Ma ± 2 sigma)	$^{207}Pb/^{206}Pb$	$^{207}Pb/^{206}Pb$
Fr86-25	Frontier granite	2340 ±8	0.1464	0.0531
		1988 ±54	0.1222	0.0931
		2928 ±20	0.2130	0.1412
		2507 ±64	0.1649	0.2057
Fr86-1	Frontier quartzite	2423 ±16	0.1568	0.2102
		2378 ±22	0.1528	0.1450
		2427 ±32	0.1574	0.0060
		2380 ±16	0.1530	0.0691
At86-64	Athona granite	2170 ±12	0.1355	0.1189
		2094 ±36	0.1298	0.1073
		1952 ±22	0.1197	0.1166
		2090 ±40	0.1295	0.1182
Bx86-66	Box granite	2064 ±24	0.1275	0.1401
		2118 ±120	0.1315	0.1530
Bx86-4	Box old granite	2920 ±12	0.2096	0.1365
		2908 ±4	0.2104	0.1180
		2779 ±40	0.1944	0.0780
Bx86-28	Box old granite	2840 ±20	0.2017	0.0865
		2924 ±36	0.2125	0.1396
		2964 ±4	0.2177	0.1422

source terrane (Table 1 and Figure 1) and the third group of the zircons from the Box and Athona mine granites, and two zircons from the Frontier mine granite, which have much younger ages of 1950 to 2100 Ma.

3. Discussion

a) Age of the Murmac Bay Group

The age of the Murmac Bay group must be younger than the age of the inherited zircons in the Frontier quartzite (ca. 2400 Ma), and it must be older than the mine granites (ca. 1950 Ma). Therefore, the Murmac Bay group is Early Proterozoic in age.

b) Age of the Mine Granites

Although the $^{207}Pb/^{206}Pb$ ages for the mine granite zircons span an interval from 2200 to 1950 Ma, the age of the mine granites must be ca. 1950 Ma for the following reasons: 1) the closure temperature for zircons is near 700°C and metamorphic assemblages show that the Goldfields area has not experienced these temperatures since the formation of the mine granites (Sibbald, 1984); 2) the $^{207}Pb/^{206}Pb$ ages obtained from lower temperature steps, which represent the outer portions of the grains of zircons from the Frontier quartzite, overlap with the older $^{207}Pb/^{206}Pb$ ages for the Athona mine granite. As field relations and the spread in ages from the zircons in the Frontier mine granite indicate that it obviously contains inherited zircons that have been partly reset during "granitization" of the Frontier quartzite, it is likely that the older zircons in the Box and Athona mine granites are also inherited zircons that have been reset, and incorporated into the mine granites. Thus, there is a complete spectrum of possible ages between the protolith zircon and the resulting event. Only the end members of the spectrum are significant. The maximum age of the mine granites is taken as the age of the youngest zircon, which is 1952 ±22 Ma.

c) Age of the 'Older' Granites

Field observations allow three age interpretations for the 'older' granites:

1) They are basement to the Murmac Bay group (cf. Lodge Bay granite) and were foliated during the first deformation and subsequently recrystallized during peak metamorphism (post D_1, pre D_2)

2) They were intruded into or replaced the Murmac Bay Group prior to the first deformation, were foliated

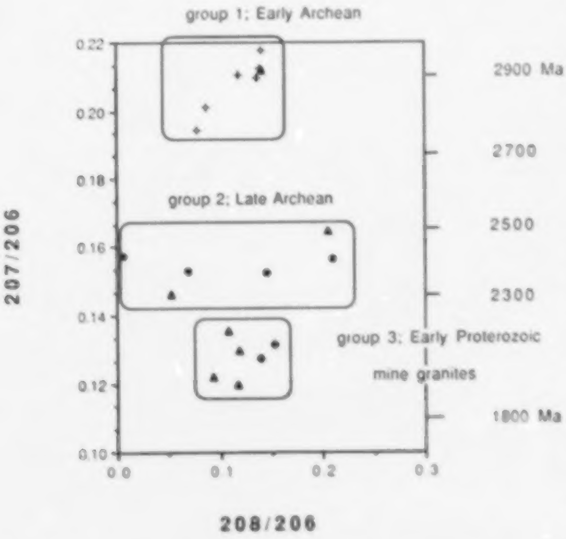


Figure 2 - $^{207}Pb/^{206}Pb$ and $^{208}Pb/^{206}Pb$ data for the zircons analyzed in this study. The ages (in Ma) corresponding to the ratios are shown on the right-hand side of the diagram. Zircons from the 'older' granites are shown as crosses; those from the Frontier quartzite as solid diamonds; Box mine granite zircons as open diamonds; Frontier mine granite zircons as solid triangles and Athona mine granite zircons as open triangles.

during the first deformation and subsequently recrystallized during peak metamorphism (post D_1, pre D_2)

3) They are metasomatic replacements of the Murmac Bay Group formed contemporaneously with the 'North Shore Plutons' during the period of peak metamorphism; first deformation foliations are 'inherited' from metasomatically altered supracrustals except in those rocks most completely granitized or in intrusive mobilizate.

The zircon data from the 'older' granites is equivocal in determining the origin of these rocks. When viewed in the context of the other zircon data and detailed field relationships, the authors envisage the 2964-2779 Ma zircon population as a detrital population, derived from 3.0 Ga basement and variably reset by magmatic, metamorphic and metasomatic events at 1950 Ma.

d) Implications for Rb-Sr Whole-rock and U-Pb Zircon Studies of the Beaverlodge Area

U-Pb zircon studies in the Beaverlodge area produced a diversity of ages (Table 1). The single-zircon Pb-evaporation study reported here has provided some insight into how the U-Pb age spectrum was produced. The difference in U-Pb ages in zircons from the mine granites that, from field relations appear to be coeval, is produced by mixing zircons formed contemporaneously with the granites with older zircons that were inherited from the Murmac Bay Group protolith. The resulting errorchrons show a low probability of fit (Persons, 1988), reflecting this mixing. This problem affects not only the mine granites but 'older' granites too, which may also contain zircons inherited from older Archean basement. Thus, ages obtained from U-Pb studies of Beaverlodge samples are potentially suspect. As zircons are clearly inherited in the mine granites, it is also possible that other minerals are inherited, so that Rb-Sr whole-rock data are also suspect. This conclusion may explain the discrepancies among the Rb-Sr ages for the hypersthene and garnetiferous felsic gneisses in the Nevins Lake Block (Table 1).

4. Conclusion

The $^{207}Pb/^{206}Pb$ ages obtained from zircons in this study are consistent with a relatively simple history for the Goldfields area. The Early Proterozoic, Murmac Bay Group basin collected sediment from ca. 3000 Ma and 2500 Ma Archean terranes. The sediments and the basement were metamorphosed to amphibolite grade at ca. 1950 Ma at temperatures sufficient to generate S-type granite melts at deeper crustal levels. The melts moved upward intruding and metasomatizing overlying Murmac Bay supracrustals. Detrital zircons from the supracrustals were variably reset, producing a spectrum of apparent ages, but with a prominence of ages that reflect the zircons in the protolith. A new population of zircons formed in melt fractions.

The 1950 Ma age for peak metamorphism and granite generation is consistent with the age of the Taltson orogen (Hoffman, 1988) and suggests that the Goldfields area was in the tectonic hinterland of this orogen.

5. Acknowledgments

D.S.O. thanks Matt Rees for discussing his sampling strategy with him and for helping him pick out the appropriate samples for dating. He also thanks Kevin Ansdell for showing him how to use the single zircon Pb-evaporation technique.

6. References

Ansdell, K.M. and Kyser, T.K. (1990): Age of granitoids from the Flin Flon domain: an application of the single-zircon Pb-evaporation technique; *in* Summary of Investigations 1990, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 90-4, p136-142.

Bell, K. and Blenkinsop, J. (1984): Saskatchewan Shield Rb-Sr geochronology: *in* Summary of Investigations 1984, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 90-4, p80.

Bikerman, M., Bell, K. and Blenkinsop, J. (1990): Rb-Sr geochronology of the Lodge Bay granite, a ca. 3.0 Ga basement in the Beaverlodge area, Saskatchewan; *in* Summary of Investigations 1990, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 90-4, p143-145.

Hoffman, P.F. (1988): United plates of America, the birth of a craton: early Proterozoic assembly and growth of Laurentia; Ann. Rev. Earth Planet. Sci., v16, p543-603.

Quirt, D. and Rees, M. (1987): Current research in mineral deposits at the Saskatchewan Research Council; *in* Summary of Investigations 1987, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 87-4, p106-163.

Persons, S.S. (1988): U-Pb geochronology of Precambrian rocks in the Beaverlodge area, northwestern Saskatchewan; unpubl. M.Sc. Thesis, Univ. Kansas. 72p.

Roberts, R.G. (1990): Structural controls of the Box and Athona deposits, Goldfields area, Saskatchewan: Sask. Energy Mines, Open File Rep. 90-2, 17p.

Sibbald, T.I.I. (1984): Gold metallogenic studies, Goldfields area; *in* Summary of Investigations 1984, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 84-4, p116-121.

Sibbald, T.I.I. and Lewry, J.F. (1980): Uranium metallogenic studies: Lodge Bay area, Lake Athabasca; *in* Summary of Investigations 1980, Saskatchewan Geological Survey, Sask. Miner. Resour., Misc. Rep. 80-4, p44-48.

Tremblay, L.P. (1972): Geology of the Beaverlodge mining area, Saskatchewan; Geol. Surv. Can., Mem. 367, 265p.

Tremblay, L.P., Loveridge, W.D. and Sullivan, R.W. (1981): U-Pb ages of zircons from the Foot Bay gneiss and the Donaldson Lake gneiss, Beaverlodge area, northern Saskatchewan; *in* Current Research, Part C, Geol. Surv. Can., Pap. 81-1C, p123-126.

Van Schmus, W.R., Persons, S.S., Macdonald, R. and Sibbald, T.I.I. (1986): Preliminary results from U-Pb zircon geochronology of the Uranium City Region; *in* Summary of Investigations 1986, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 86-4, p108-111.

Deformation in the Star Lake-McLennan Lake Area, La Ronge Domain[1]

Jie He[2] and R.G. Roberts[2]

Jie, H. and Roberts, R.G. (1991): Deformation in the Star Lake-McLennan Lake area, La Ronge Domain; in Summary of Investigations 1991, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 91-4.

The project is a continuation of the 1990-91 study of the history of deformation of the Star Lake and Island Lake plutons of the La Ronge Domain (Roberts, 1990). The objective of that study was the development of a structural model for the formation of the gold deposits in the plutons (Star Lake, Jolu and Jasper mines, and the Roxy showing) in the context of the metamorphic and structural evolution of the host granitic plutons. This project extends this study to the regional setting, with the objective of placing the evolution of the hydrothermal fluids, including the gold-bearing fluids into the context of the metamorphic and structural evolution of the La Ronge Domain.

Geological mapping in 1991, on the scale of 1:20 000, examined the structural aspects of the McLennan Lake Tectonic Zone (Lewry, 1983), and the volcanic and sedimentary rocks between the Alpat Lake Tectonic Zone (Thomas, 1985) and the Star Lake and Island Lake plutons.

1. General Geology

The study area (Figure 1) is underlain by mafic to felsic metavolcanic and metavolcaniclastic rocks of the Central Metavolcanic Belt of the La Ronge Domain. The supracrustals are intruded by a suite of acid to basic igneous rocks, which includes the Star Lake and Island Lake plutons. The area is bounded on the southeast by the McLennan Lake Tectonic Zone, which separates the metavolcanic rocks from the metasediments of the McLennan Group (Lewry 1983; Thomas 1985). The rocks have been metamorphosed to the lower amphibolite facies.

2. Structural Geology

Figure 1 summarizes the variation in the strike of the foliation in the metasedimentary rocks of the McLennan Group, the metavolcanic and associated metasedimentary rocks, and the southern part of the Island Lake pluton. The trend shown in the figure is that of the dominant or principal foliation observed in the outcrops. In many outcrops there is a single foliation; this is particularly so in the high strain sections of the McLennan Lake and Alpat Lake Tectonic Zones. However, elsewhere, two or more foliations may occur, and these tend to curve or deflect into local high strain-zones, a few centimetres wide. Figure 1 implies continuity of the trend of the foliation with changing attitude. This can be demonstrated in the south and west part of the map area, but in the north there is insufficient outcrop to be certain that the trend of the foliation varies continuously.

The deformed rocks, are characterized by a steeply plunging mineral lineation defined by the preferred orientation of feldspar, hornblende and biotite. The lineation is intensely developed in shear zones, and less developed, but still prominent in foliated rocks outside the shear zone. The contoured stereographic plots (Figures 2, 3 and 4), illustrate the relationship between the attitude of the lineations, and the attitude of the foliations on which they lie.

a) McLennan Lake Tectonic Zone

The intensity of the strain in the McLennan Lake Tectonic Zone increases towards the contact of the metavolcanic rocks with the sedimentary rocks of the McLennan Group. In the sedimentary rocks of the McLennan Group, only a single foliation with the associated lineation is generally developed (Figure 2), and there is little direct evidence for layer parallel shearing. The sense of movement on the shear zone was obtained in lower strained rocks where the foliation is deflected into narrow zones of higher strain. The movement on the ductile shear is northwest side down.

The early foliation is locally deformed to form rotated boudinage structures, and mesoscopic scale folds, that indicate a dextral movement. The fold axes have variable plunges to the northeast indicating oblique slip (Figure 5). Striations, developed locally on shear surfaces plunge to the southwest are believed to be associated with this stage of deformation.

b) Alpat Lake Tectonic Zone

The Alpat Lake Tectonic Zone (Thomas, 1985) trends north-northeast and is defined by strongly deformed fine grained phyllonites interlayered with foliated metasedimentary rocks and minor metavolcanic rocks. Because of the fine grain size of the rock, it is difficult to make good measurements of lineations. The few reliable data indicate a steep plunge to the north. Sinistral movement is interpreted from the relationships between foliations.

---

(1) Saskatchewan Project A.142, from which this report derives, was funded in 1991 under the Canada-Saskatchewan Partnership Agreement on Mineral Development 1990-95.

(2) Department of Earth Sciences, University of Waterloo, Waterloo, Ontario, N2L 3G1.

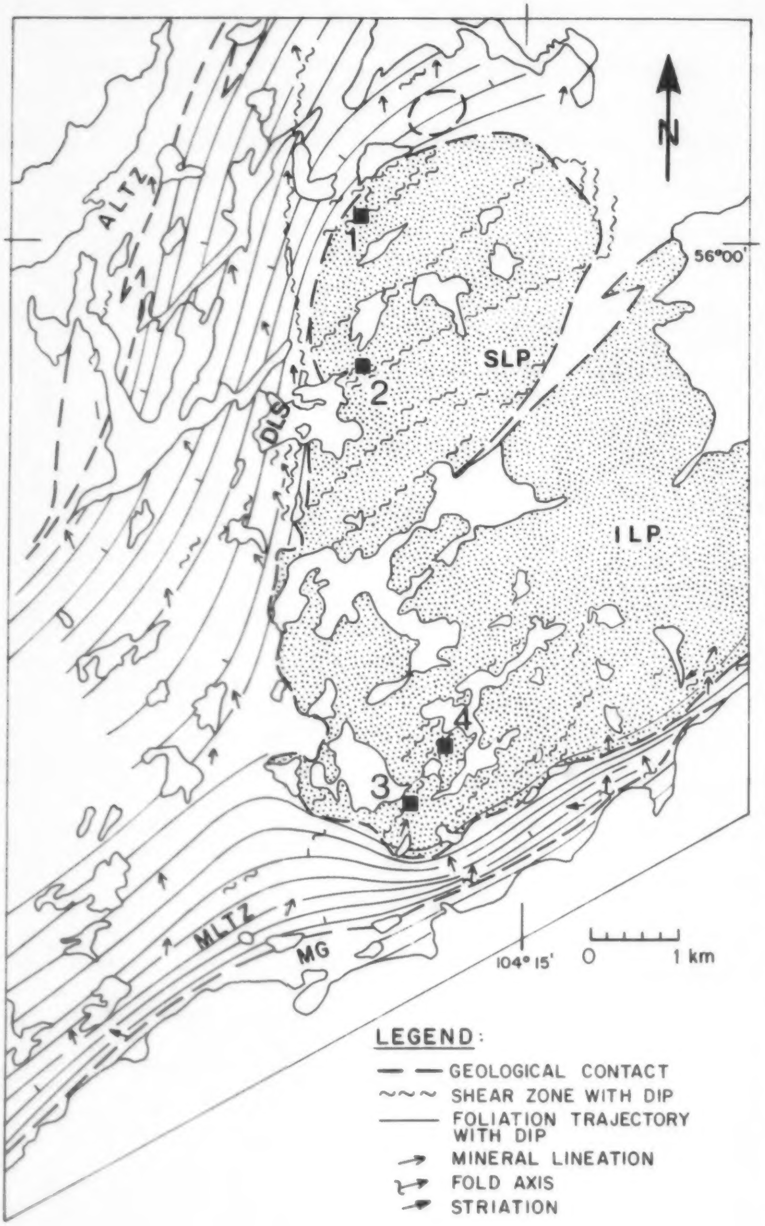


Figure 1 - The Star Lake-McLennan Lake area showing the trend of the foliations. ILP, Island Lake pluton; SLP, Star Lake pluton; MLTZ, McLennan Lake Tectonic Zone; ALTZ, Alpat Lake Tectonic Zone; DLS, David Lake shear; MG, McLennan Group. Gold deposits: 1, Jolu mine; 2, Star Lake mine; 3, Jasper mine, 4, Roxy showing.

c) David Lake Shear

The David Lake Shear (Poulsen *et al.*, 1987) is a north-trending structure in the metavolcanics to the west of the Star Lake pluton. The structure forms a well defined lineament on air photographs, but there are very few exposures. Shears associated with the structure strike between 355° and 020°, and dip steeply to the west. Mineral lineations plunge steeply to the north-northwest, and sinistral movement is indicated by the deflection of foliations into high strain zones.

3. Discussion

The observations on the McLennan Lake Tectonic Zone support the conclusions of Poulsen *et al.*, (1987) that the zone has undergone two stages of deformation: an early stage of dip slip ductile shearing followed by oblique slip, dextral movements. These two stages of deformation are also found in the gold-bearing shear zones of the Island Lake and Star Lake plutons. The principal (most extensive) gold-bearing quartz veins are associated with the early dip slip movements. In the Island Lake pluton, the second stage is associated with brittle-ductile deformation which resulted in folding of the shear fabric and the formation of gold-bearing quartz veins, of limited strike length, trending approximately north-south. The mesoscopic scale folds and associated veins are developed locally in the shears of the pluton, but are abundant in the Jasper and Roxy deposits. The orebodies in the Jasper deposit plunge steeply, parallel to the early lineation, and are unaffected by the second stage of deformation. At the Jolu mine in the Star Lake pluton, the second stage of deformation is ductile, and the axial surfaces of the mesoscopic folds are parallel or sub-parallel to the folded shear fabric. The sense of movement is dextral, but the folds plunge to the southwest. At the mesoscopic scale, the quartz veins which predate this stage of deformation are deformed to form lenses which plunge with the axes of the folds. The characteristic southwest plunge to the Jolu and Star Lake orebodies is probably the result of disruption of the earlier veins by this second stage of deformation.

The steeply plunging mineral lineation is characteristic of deformed rocks both within recognized shear zones, where it is most strongly developed, and in foliated rocks outside the shear zones. The variation of the plunge of the lineation depends on the attitude of the foliation with which it is associated. These same characteristics may be seen at the mesoscopic scale, where

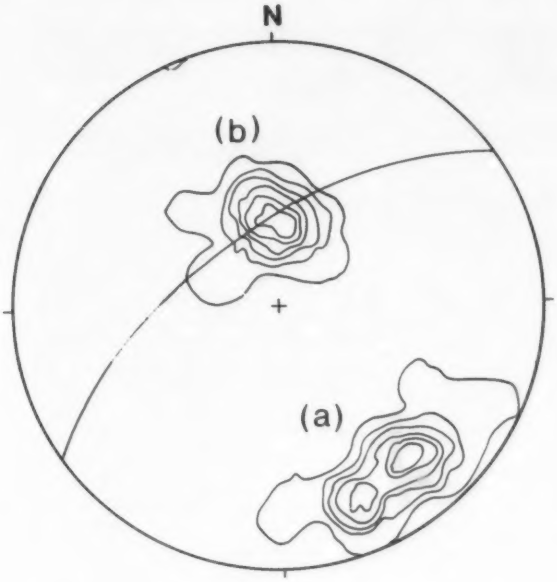


Figure 2 - Stereographic plot of foliations and lineations from the McLennan Lake Tectonic Zone. (a) Poles to the foliations, N=52. (b) Lineations measured on the foliations, N=52. Contour interval is 4% per 1% area. The mean foliation is shown as a plane.

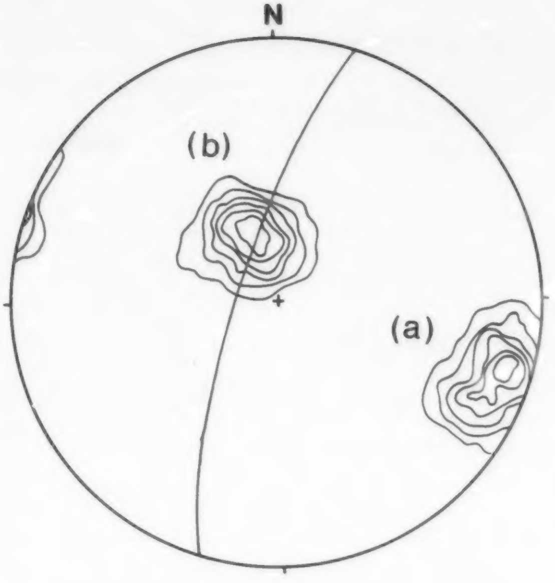


Figure 3 - Stereographic plot of foliations and lineations from the metavolcanic and metavolcaniclastic rocks. (a) Poles to the foliations, N=41. (b) Lineations measured on the foliations, N=41. Contour interval is 4% per 1% area. The mean foliation is shown as a plane.

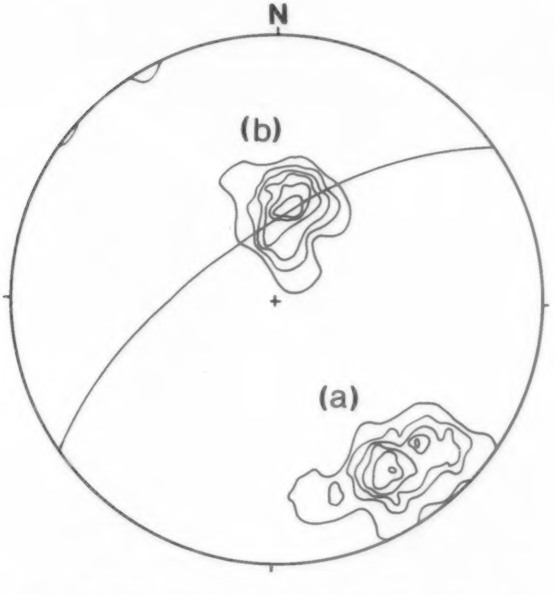


Figure 4 - Stereographic plot of foliations and lineations from the metavolcanic and metavolcaniclastic rocks. (a) Poles to the foliations, N=21. (b) Lineations measured on the foliations, N=21. Contour interval is 4% per 1% area. The mean foliation is shown as a plane.

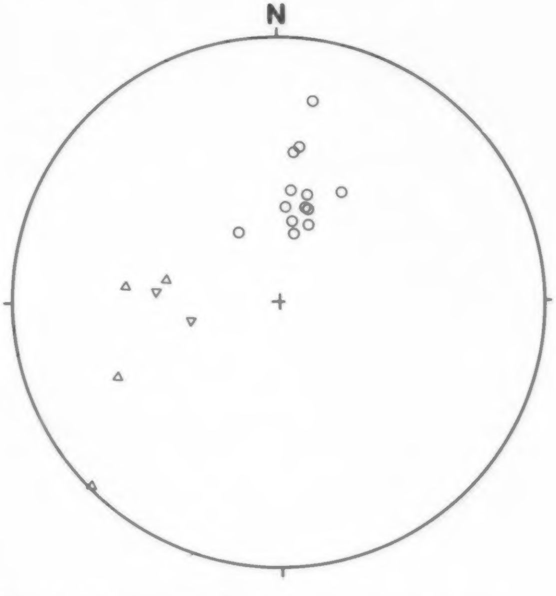


Figure 5 - Stereographic plot of fold axes (circles) and striations (triangles) from the McLennan Lake Tectonic Zone.

the regional foliation is deflected into local high strain zones. Thus the fabrics suggest that the various shear structures were developed during a single period of deformation.

4. References

Lewry, J.F. (1983): Character and structural relations of the "McLennan Group" meta-arkose, McLennan-Jaysmith Lakes area; *in* Summary of Investigations 1983, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 83-4, p49-55.

Poulsen, K.H., Ames, D.E. and Galley, A.G. (1987): Structural studies in the northern part of the La Ronge Domain; *in* Summary of Investigations 1987, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 87-4, p107-114.

Roberts, R.G. (1990): Structural controls on gold deposits in the Star Lake and Island Lake Plutons; *in* Summary of Investigations 1990, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 90-4, p55-59.

Thomas, D.J. (1985): Geology of an area around Star Lake; Sask. Energy Mines, Open File Rep. 85-6, 83p.

Alteration Halos Associated with Gold-bearing Shear Zones in the Star Lake and Island Lake Plutons[1]

Edward C. Appleyard[2]

Appleyard, E.C. (1991): Alteration halos associated with gold-bearing shear zones in the Star Lake and Island Lake Plutons; *in* Summary of Investigations 1991, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 91-4.

Gold-bearing shear zones in the Star Lake and Island Lake Plutons, located about 150 km northeast of La Ronge, have been known since the 1940s (Murphy, 1986) but were developed and successfully exploited only in the late 1980s. The two plutons which host the majority of discoveries in this area consist of adjacent, roughly elliptical bodies with the Star Lake Pluton situated north of and in faulted contact with the Island Lake Pluton. This report provides some initial results of a detailed study of the geochemistry and petrology of rocks associated with and involved in the events which resulted in these structures and their mineralization.

1. Geological Setting

The Star Lake Pluton is marked by a strong compositional zonation from a gabbro-diorite marginal phase, through monzonite-granodiorite-quartz monzonite "transitional" phases to a central granitic (*sensu stricto*) core (Thomas, 1984, 1987; Poulsen *et al.*, 1986). The mineralized shear zones strike *ca.* 060°, dip steeply and have strike lengths of 1 to 4 km. Included in this group are the shears which host the Mallard Lake (Decade), Rod (Jolu), Rush/Pie, 21 (Star Lake), Blindman and Tamar gold occurrences (Poulsen *et al.*, 1986).

The Island Lake Pluton is more uniform, consisting primarily of leucogranite with monzonite, quartz monzonite and granodiorite zones (Thomas, 1985). Shear directions are more variable than in the Star Lake Pluton but lie in the northeast quadrant. In the Broeder Lake area the shear systems that host the Jasper and Roxy deposits strike *ca.* 040° and 030° respectively, but the Roxy Zone comprises a network of 010° and 040° zones (Roberts, 1990).

Shear zones have widths up to 30 m (Ames *et al.*, 1987), but are more commonly much narrower. The high-strain structures pinch and swell both horizontally and vertically, bifurcate and have splays which deviate into the walls of the main vein. Ore grade mineralization is similarly irregularly distributed within individual shears and is interspersed with sub-economic or barren sections.

2. Previous Studies

It has hitherto generally been believed that hydrothermal alteration is essentially absent from the wall-rocks of the shears including, most importantly, rocks contiguous to gold-bearing shoots within shears. Poulsen *et al.* (1986), for example, stated that the mylonitic and protomylonitic rocks of the 21 Zone shear at the Star Lake Mine are chemically indistinguishable from the less deformed quartz monzonite and illustrated that the host rocks and their sheared equivalents plot congruently on a Na_2O-K_2O-CaO ternary diagram. Minor sericitization by "simple hydration of microcline" which produces a slight colour change from pinkish or grey to greenish was the only mineralogical effect noted by these authors. In another paper Poulsen (1986) compared major element analyses of a host rock from "Star Lake" and its foliated equivalent on an Al_2O_3-normalized basis and found significant differences only in the L.O.I. values.

Hematitic alteration, either confined to specific fracture sets or more penetratively developed in granite adjacent to fractures, was regarded by Poulsen *et al.* (1986) to be of late origin, and unrelated to the shearing events. On the other hand, geologists at the Jasper Mine in the Island Lake Pluton indicated that they believed that the "granite" host rocks are *oxidized* and characterized by magnetite as an accessory mineral but that the shear zones are *reduced* and characterized by pyrite (Poulsen, pers. comm., 1990). They had tried to exploit this difference by using magnetic susceptibility as a prospecting technique but found the contrast too low and the shears too narrow for the method.

3. Objectives of Present Study

The present study was designed to test the hypothesis that the hydrothermal event which led to the introduction of gold into the sheared structures must also have had other geochemical and mineralogical expressions and, if so, to establish criteria to assist explorationists to recognize "prospective" alteration effects.

---

(1) Saskatchewan Project A.143, from which this report derives, was funded in 1991 under the Canada-Saskatchewan Partnership Agreement on Mineral Development 1990-95.

(2) Department of Earth Sciences, University of Waterloo, Waterloo, Ontario, N2L 3G1

4. Study Plan

The region was visited August 24 to September 7,1990. Well-exposed, stripped and/or trenched areas of six shear zones were selected for detailed sampling. These included:

1) a trenched zone on the shore of David Lake exposing a shear peripheral and to the northwest of the 21 Zone on the Star Lake Mine property,

2) a stripped portion of the Rush Lake Shear Zone overlooking the southwest end of Rush Lake,

3) the extensively stripped exposures of the Roxy Zone between the Jasper Mine and Broeder Lake,

4) a trench crossing the exposed Jasper Shear Zone on the Jasper Mine property,

5) the extensively stripped James Zone immediately to the southwest of the Jasper Zone, and

6) a stripped outcrop near the southwesterly limit of the Rod Zone on the Jolu Mine property.

This report presents data and initial interpretations relating to the first five of the above occurrences.

One hundred and fifty-one samples were collected from these localities of which 93 were analyzed by Activation Laboratories Ltd. for 54 components including Fe_2O_3 and FeO. Densities of all specimens were measured using a Beckman Model 930 Air Comparison Pycnometer.

5. Mass Balance Corrections

Mass balance corrections for changes in density and volume during the shearing and subsequent hydrothermal alteration were carried out using Gresens (1967) general metasomatic equation and procedures outlined by Appleyard (1990). Reference "parental" samples showing no evidence of alteration and minimal effects of shearing were collected as close to sampling sites as possible.

Histogram plots of "zero change volume factors" identified a relatively consistent group of elements as having predominantly immobile characteristics during the alteration and some or all of these elements were therefore used to calculate volume change factors (FV values) for the strained and altered rocks. The immobile element suite includes Ti, Al, Sc, Y, Zr, Nb, Yb, Lu, and Hf. All nine elements were used to constrain the volume change relationships of the James, Jasper and Star Lake suites. Ti, Al, Sc and Nb were utilized for the Rush Lake occurrence and Ti, Al, and Nb in the case of the Roxy occurrence. A few samples in most suites contain some otherwise apparently immobile elements displaying outlying or incongruent behaviour. When mean volume change factors were being calculated outlying data that fell >1 standard deviation on either side of the original mean were excluded and the thereby restricted mean recalculated.

a) Mass Changes

Volume change parameters for the sheared and altered rocks fall between 0.859 (a 14 percent volume loss) and 3.28 (a 228 percent volume increase). Patterns of volume change as a function of distance from median lines of shear zones are illustrated in Figure 1. The general pattern is for small volume gains or losses to have occurred in the foliated rocks flanking the shear. Close to the highest strain zone volume changes may become erratic and of larger magnitude (gains and losses) but within the main shear zone the majority of the

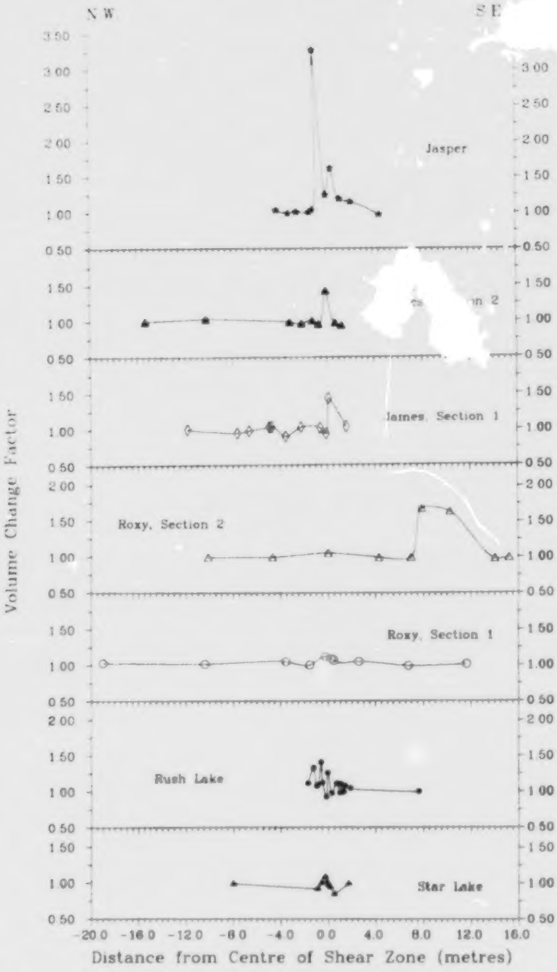


Figure 1 - Variations in calculated volume change factors (FV values) along seven cross-sections of shear zones in the Star Lake and Island Lake Plutons. The volume change factor represents the ratio of the volume of the rock after alteration to that of a reference mass of the rock before alteration. An FV value of 1.0 represents no volume change; values <1.0 represent volume loss and >1.0, volume increase.

values indicate strongly additive effects. In most of the rocks involved in this study density changes were slight so volume change factors are almost identical to mass change factors.

6. Classification of Rock-types

For purposes of this study the textural classification devised by Poulsen *et al.* (1986), augmented with some mineralogical descriptors to draw attention to changes in modal compositions, provided a suitable basis for discriminating various strain facies. This classification recognizes four principal states of strain:

1) **Least deformed** - lacking megascopic signs of foliation, although in thin section some preferred orientation of biotite may be evident as well as strain in quartz and very minor grain boundary recrystallization;

2) **Weakly foliated** - a readily visible megascopic and microscopic foliation, but porphyritic minerals are not appreciably disrupted;

3) **Protomylonitic** - strong undulose foliation, porphyritic minerals commonly strongly strained and abraded but still recognizably megacrystic, <10 percent of granulated matrix; and

4) **Mylonitic** - very strongly foliated and layered rock in which >50 percent of the original grains have been reduced to a fine-grained matrix.

A very significant amount of strain recovery recrystallization has affected most samples since shearing occurred so that in thin sections the more highly strained and comminuted grains are extensively replaced by polygonal, relatively low-strain grains and it is primarily the megacrystic grains that preserve undulose and granulated textures. For this reason it seems appropriate to use the terminology for partially and extensively recrystallized tectonites outlined by Spry (1969) whereby rocks with the texture described above are termed **hartschiefer** and mylonitic facies with extensive recrystallization are termed **blastomylonite**.

Mineralogical changes associated with increasing evidence of strain and/or proximity to shear zones depend somewhat on the nature of the host-rock. Amphibole-bearing host rocks tend to become enriched in biotite. Minor carbonate may appear. Biotite-bearing rocks tend to acquire discrete flakes of muscovite. In some samples plagioclase may become somewhat turbid and saussuritic but the alteration is generally not characterized by the development of fine-grained, retrogressive-looking effects.

In some microcline-bearing facies, careful observation reveals the presence of a second, younger generation of microcline and the modal abundance of microcline increases, typically in association with the appearance of significant muscovite. This effect is particularly noteworthy in rocks from the Roxy Zone. Rims of albite and/or myrmekite have been reported by Thomas (1985) on plagioclase/microcline mutual boundaries but these are indistinguishable from normal retrogressive intergranular exsolution rims and are not believed to be related to shear-focused hydrothermal alteration.

Cross-section 1 from the James Zone is noteworthy for the abundance of epidote across much of the outcrop and by extensive fracture-coating and penetrative hematite.

These mineralogical changes correlate strongly with the textural classification outlined above and with the proximity to shear zones; they clearly indicate that the alteration which was focused on the shear zones took place under conditions of temperature and pressure and fluid composition appropriate to lower amphibolite facies conditions.

7. Lithogeochemistry

An indication of the nature of some of the chemical effects of alteration is provided in Table 1 and Figures 2 to 5.

a) Roxy Zone Compositional Changes

In Table 1 average analyses for least altered and four facies of sheared granodiorite from the Roxy Zone are listed according to the rocks' lithological classifications. The least altered rock is initially biotite-bearing but as the textural evidence of strain increases muscovite and then second generation microcline appear permitting the discrimination of several facies of hartscheifer. The most intensely sheared rock comprises very strongly foliated and banded phyllonitic rock extensively permeated and veined with cherty to glassy quartz veins.

Mean volume change factors (FV) increase gradually with the intensity of strain until the mylonitic zone is reached where a sudden increase is recorded. These values reflect the fact that dilatation accompanies cataclasis. Fluids attracted to the lower pressure and higher permeability zones cause replacement, deposition and solution effects to be focused there. To some extent, and notably in some samples, the volume increase effect is countered by the dissolution of intensely strained tectosilicate minerals, especially quartz, but at least in the Roxy Zone the net effect is of progressively increasing volume.

Si and O increase in the largest proportions from the least altered granodiorite to the granodiorite blastomylonite. The O:Si molar proportion is 1.97 indicating that a tectosilicate is the predominant mineral form created and modal studies indicate it is predominantly quartz. Total iron has a pattern that is found in a number of elements; modest losses appear in the slightly to moderately strained facies and gains are recorded in the most strongly deformed rocks. Mg, Ca and Na all exhibit strong systematic losses throughout most of the sequence with the exception that slight upturns in the Ca and Na trends occur in the blastomylonite facies. These changes reflect destruction of plagioclase and biotite and their replacement by microcline and muscovite

Table 1 - Average Analyses of Lithological Facies from the Roxy Shear Zone. Values are **net compositions**, *(i.e. are corrected for volume and density changes during alteration) and represent the amount of each element remaining after 100 g (1 tonne for trace elements) of the host rock have been altered. Major element values are arithmetic averages; trace element values are logarithmic averages. The values for Oxygen are calculated. Rock-type mnemonics are: Gd, least altered granodiorite; Bgdh, biotite granodiorite hartschiefer; Mgdh, muscovite-biotite granodiorite hartschiefer; Mgdh+, microcline-muscovite granodiorite hartschiefer; Ubmy, granodiorite blasto-ultramylonite. Additional elements analyzed but not included here because most values fall below the limit of detection are Be, Se, Br, Cd, Te, Ta, W, Hg, Tl and Bi.*

Unit	Gd	Bgdh	Mgdh	Mgdh+	Ubmy
N	4	7	3	4	2
S.G.	2.660	2.658	2.655	2.652	2.659
FV	1.000	1.003	1.020	1.069	1.619
Si	32.9	33.1	33.5	36.3	60.7
Ti	0.138	0.137	0.136	0.139	0.131
Al	7.88	7.91	7.93	7.79	7.72
Fe^{3+}	0.56	0.64	0.48	0.69	1.06
Fe^{2+}	0.95	0.88	0.90	0.77	0.76
TFe	1.51	1.53	1.37	1.46	1.81
Mn	0.029	0.032	0.040	0.025	0.025
Mg	0.63	0.64	0.60	0.57	0.50
Ca	1.61	1.44	1.40	1.09	1.25
Na	3.18	3.04	2.65	2.52	2.56
K	3.00	3.22	3.48	4.23	3.03
P	0.035	0.037	0.033	0.033	0.007
$O^{calc.}$	47.9	48.1	48.5	51.3	79.1
S	0.023	0.030	0.113	0.202	0.097
Sc	2.6	2.5	2.6	2.7	3.2
V	24	23	22	25	4
Cr	13	16	22	17	15
Co	4.4	3.8	4.0	4.7	4.5
Ni	7	10	10	11	16
Cu	2.0	4.3	5.5	9.4	27.3
Zn	18.8	19.8	21.2	25.8	32.0
Ga	3.7	3.8	3.3	3.3	3.1
As	0.7	1.0	1.3	1.8	2.3
Rb	97	100	98	115	99
Sr	532	447	357	338	232
Y	6.2	6.3	6.1	5.6	2.8
Zr	122	100	113	118	103
Nb	10.7	10.9	10.4	11.1	13.7
Mo	0.27	0.25	0.37	0.99	6.6
Ag	0.05	0.09	0.22	0.39	0.95
Cs	1.2	1.4	1.1	1.4	2.0
Ba	1060	1021	1094	1277	1203
La	14.6	16.5	16.2	16.3	5.7
Ce	28.2	30.1	30.5	33.2	20.0
Nd	9.2	9.5	10.5	10.2	4.0
Sm	1.4	1.4	1.5	1.4	0.72
Eu	0.4	0.4	0.4	0.4	0.3
Yb	0.39	0.41	0.40	0.43	0.32
Lu	0.07	0.08	0.08	0.07	0.02
Hf	2.5	2.6	2.4	2.8	3.5
Au	0.004	0.019	0.079	0.436	8.60
Pb	3.0	3.8	4.3	8.7	11.2
Th	5.5	6.1	6.6	5.5	7.8
U	1.9	1.3	2.5	2.8	5.8

respectively. K is notable as being very significantly introduced throughout the sequence with the exception of the quartz-flooded blastomylonite. S increases throughout the series, except for a downturn in the blastomylonite. These exceptions in the blastomylonite may be a function of the limited sample set of this unit from the Roxy Zone; they are not characteristic of all the data sets collected.

Among the trace elements the significant introductions of the base metals Cu, Zn, Mo and Pb as well as the modest gains in Th and stronger introduction of U should all be noted. Similarly, as one would anticipate, there are very strong progressive increases in both Au and Ag. Among the lesser trace elements, gains are recorded by As and probably Cs and Ba. In the case of the rare earth elements it appears that the light REE increase slightly while the heavy REE remain constant with the exception of the blastomylonite facies where all have been strongly depleted. The relative immobility of the heavy REE, Yb and Lu, was recognized previously by their inclusion amongst the elements identified as immobile in some of the other suites.

Sb, Hg and Bi probably also should be counted among the elements that were introduced. The highly strained facies contain levels of these elements that are above the sensitivity limit of analysis but the least and slightly deformed rocks contain unmeasurably low levels so the data have not been included in Table 1.

Sr is exceptional in the degree to which it has been removed. This is probably a reflection of the destruction of plagioclase. Gallium appears to display progressive losses as well.

Similar gain-loss systematics can be observed in the other suites sampled. It is clear that the shear zones and their wall-rocks have undergone significant and systematic compositional changes during the introduction of hydrothermal fluids that flowed through the shears and were responsible for the introduction of the precious metals. It is probable that these changes were not recognized by previous workers because they were using raw analytical data where the patterns of fluxes were obscured by the uncorrected "noise" of the volume changes.

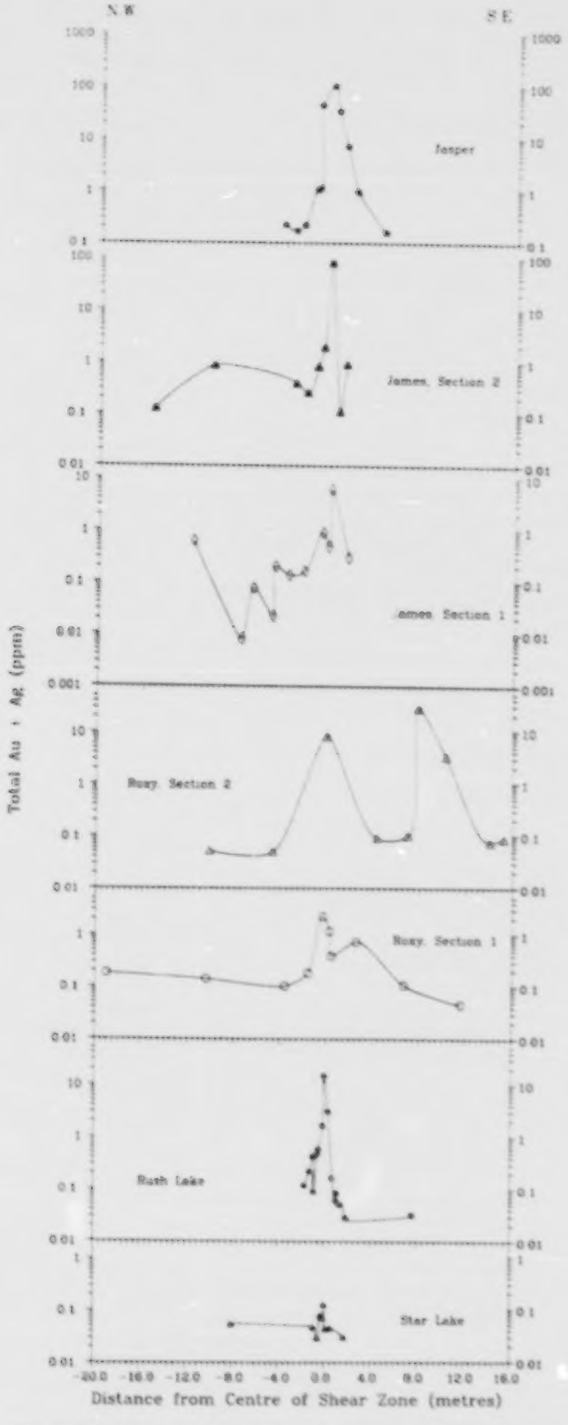


Figure 2 - Variations of total Au + Ag represented as "net composition" values, i.e. corrected for mass changes during alteration, and expressed as ppm with respect to shear zones. More information on the calculation of **net compositions** *can be obtained in Appleyard (1990).*

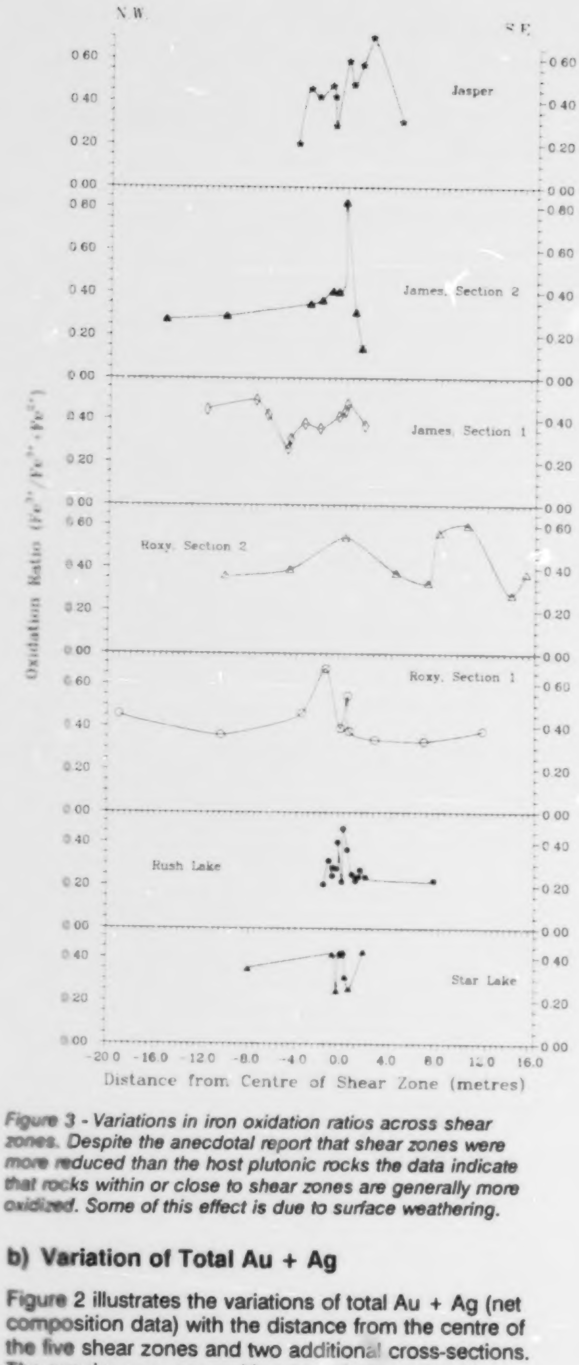


Figure 3 - Variations in iron oxidation ratios across shear zones. Despite the anecdotal report that shear zones were more reduced than the host plutonic rocks the data indicate that rocks within or close to shear zones are generally more oxidized. Some of this effect is due to surface weathering.

b) Variation of Total Au + Ag

Figure 2 illustrates the variations of total Au + Ag (net composition data) with the distance from the centre of the five shear zones and two additional cross-sections. The graphs are arranged in approximate order of decreasing precious metals from top to bottom. It is apparent from the shapes of some of the curves that many of the shear zones are markedly asymmetrical. In fact, it is commonly the case that one margin of a shear may be relatively sharp while the other margin is gradational. The Roxy, Section 2 crossed two shear zones which are marked by very prominent Au + Ag peaks whereas because of the diverging trends of the two shears only one was crossed in Roxy, Section 1. In some cases the increases in Au and Ag above background (about 0.05 ppm) can be detected over 10 m from the shear zone but in other cases the peaks extend only 1 or 2 m from the shears.

c) Oxidation Ratios ($Fe^{3+}/Fe^{3+}+Fe^{2+}$)

Since changes of oxidation ratios had been suggested as a signature of shear zones it seemed apposite to examine their patterns in the current seven sections. The data are portrayed in Figure 3. The results suggest that the patterns are opposite to those reported, i.e. most sections are more oxidized within or adjacent to the shears than at a distance. A warning is, however, necessary inasmuch as highly fissile, phyllonitic facies within and marginal to some of the main shear zones were deeply weathered and fresh rock was unobtainable, so in some cases the oxidation is largely a result of recent weathering. This limitation does not apply to all the data so some of the oxidation must be related to the hydrothermal processes. The generally elevated values noted across the James, Section 1, reflect the notably epidote-rich and hematitic rocks found in this area of the James ridge. Both shear zones on Roxy, Section 2 are marked by prominent peaks although only the southeastern shear on Roxy, Section 2 contains notably elevated Au + Ag.

d) $K^+/(Mg^{2+}+Ca^{2+}+Na^+)$ Ratio

Amongst the major elements it was recorded above that K was notable as an addition and Mg, Ca and Na were all lost. As an index of alteration, the ratio of these groups of elements strongly portrays the hydrothermal effects. The data are illustrated in Figure 4. It is evident that all sections show peaks in this parameter closely associated with the shear zones and furthermore the height of the peaks is closely proportional to the Au + Ag levels. Note that the southeast end of the trend on Roxy, Section 1 shows a marked rise as the second shear zone (which is present on Roxy, Section 2), is approached. It appears that strongly developed, mineralized shears are marked by prominent changes in this ratio for distances of 5 to 15 m from the shear margins.

e) Base Metals ± Sr Ratios

Amongst the trace elements some of the most notable changes involved the base metals, Cu, Zn, Mo, and Pb, all of which were gained, and Sr which was lost. Within the base metal group it appeared that Cu and Pb were more strongly gained than Zn and therefore that the Cu/(Cu + Pb) ratio or the (Cu + Pb)/(Cu + Zn + Pb) ratio would also be indicators of hydrothermal alteration. Another variation on this approach was to ratio the four added base metals against Sr values which diminished. An index expressed as Log((Cu * Zn * Mo * Pb) / Sr) was therefore devised to exploit this relationship. Finally, the sum of Cu + Pb (net compositions) was investigated.

These four parameters proved to be nearly equally strong indices of the strained facies. One of these is presented here in Figure 5, *viz.* the sum of Cu + Pb.

8. Discussion

The first objective of this study was to investigate the hypothesis that the hydrothermal processes which resulted in the introduction of Au into the sheared structures would have had other geochemical expressions, i.e. that there would be wall-rock alteration halos adjacent to the veins. The limited selection of results presented above makes it evident that such halos exist and are illustrated in a wide range of elements. Furthermore, it has been demonstrated that the shear zones are also recognizable by the fact that they have resulted in dilatancy during cataclasis and this feature, plus the converse effect of tektosilicate dissolution (predominantly that of quartz) in strained facies, can be tracked through changes in the volume change factors (FV values). It is concluded that such well-preserved changes and patterns were not recognized previously due to volume changes during the shearing and hydrothermal processes which resulted in spurious changes in concentration of the elements on a sample by sample basis. Thus the advantage of undertaking mass balance corrections is demonstrated. It should be noted that the plutonic host rocks of these shears provide an ideal setting for mass balance corrections inasmuch as problems of host-rock inhomogeneity are minimized.

The second objective was to investigate criteria to identify "prospective" structures. A small selection of possible alteration indices has been presented that accomplish this goal to varying degrees. All the indices shown (Figures 2 to 5) are capable of picking out most of the shear zones studied and correlate well in magnitude with the precious metal values found there. The ratio of $K^+/(Mg^{2+}+Ca^{2+}+Na^+)$ appears to be slightly preferable to the others, insofar as, in several cases it results in broader peaks, i.e. the anomalous alteration effects can be detected at greater distances from the shear zones. Under the best conditions it appears that alteration can be detected by this approach at distances up to 15 m from the shear but in other situations, i.e. where the shear zone margin is sharp against unsheared rock, the distance of detection may be as little as one or two metres.

It must be recognized that in the examples presented here the index **Au + Ag** is essentially as good as any of the other indices. Apparently, the elements Au and Ag are as widely dispersed in the alteration zones as K, for example, and concentration factors in the anomalous zones are very high relative to the usual background levels in the host-rocks. Nevertheless, the type of alteration indices developed here can all be computed with little effort and cost, and may be more effective when used as a group rather than individually. The minimum data required would be "research quality" analytical data for the immobile elements Ti, Al, Sc, Y, Zr, and Nb and possibly Yb, Lu, and Hf, as well as for the mobile elements Mg, Ca, Na, K, Cu, Zn, Sr, Mo, Ag, Au, and Pb with total $Fe_2O_3^{total}$ and FeO as elective additional ele-

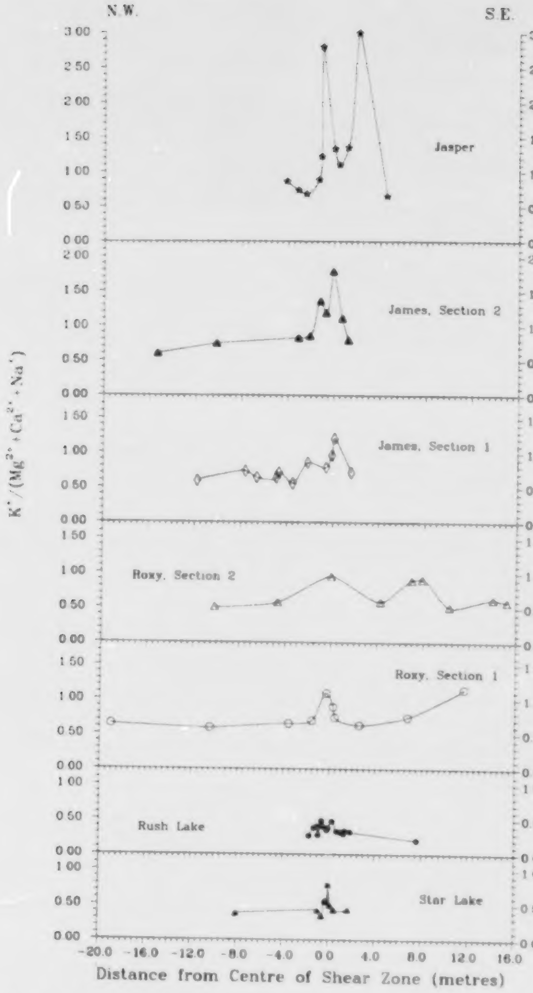


Figure 4 - Variations of the alteration index $K^+/(Mg^{2+}+Ca^{2+}+Na^+)$ with distance from shear centres.

ments. Determinations of some of the other elements identified as being mobile would permit the computation of other alteration indices. In addition, it is necessary to have density determinations measured on all samples.

Additional work is being done on the Rod Shear Zone in two locations on the Jolu Mine property during the current year and on drill core from the Muskeg Zone of the Jasper deposit. These studies will shed light on the scope of alteration in highly productive sections of the respective deposits and will explore the problems encountered when shearing has juxtaposed different lithologies across the shear zone.

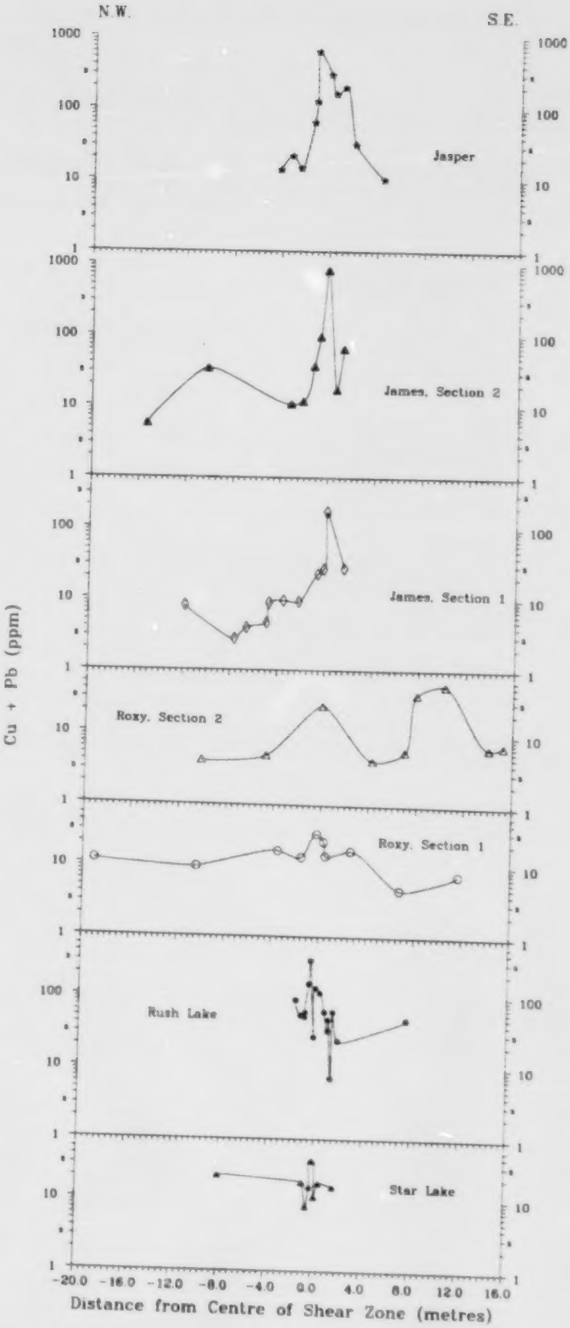


Figure 5 - Variations of total Cu + Pb with distance from centres of shear zones. The values of Cu and Pb used are "net composition" values as in Figure 3.

9. References

Ames, D.E., Poulsen, K.H. and Galley, A.G. (1987): Geology and gold deposits, Star Lake Pluton, La Ronge Domain, Saskatchewan; Geol. Surv. Can., Open File 1469, 1:10 000 scale map with marginal notes.

Appleyard, E.C. (1990): Mass balance corrections applied to lithogeochemical data in mineral exploration; *in* Beck, L.S. and Harper, C.T. (eds.), Modern Exploration Techniques, Sask. Geol. Soc., Spec. Publ. 10, p27-40.

Gresens, R.L. (1967): Composition - volume relationships of metasomatism; Chem. Geol., v2, p47-65.

Murphy, W.L. (1986): Geology and exploration history of the gold mineralization in the Star Lake area, Saskatchewan; *in* Clark, L.A. (ed.), Gold in the Western Shield, CIM Spec Vol. 38, p212-220.

Poulsen, K.H. (1986): Auriferous shear zones with examples from the western shield; *in* Clark, L.A. (ed.), Gold in the Western Shield; CIM Spec. Vol. 38, p86-103.

Poulsen K.H., Ames, D.E. and Galley, A.G. (1986): Gold mineralization in the Star lake pluton, La Ronge belt, Saskatchewan: a preliminary report; *in* Current Research, Part A, Geol. Surv. Can., Pap. 86-1A, p205-212.

Roberts, R.G. (1990): Structural controls on gold deposits in the Star Lake and Island Lake Plutons; *in* Summary of Investigations 1990, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 90-4, p58-59.

Spry, A. (1969): Metamorphic Textures; Pergamon Press, Oxford, 350p.

Thomas, D.J. (1984): Geological mapping, Star Lake area (part of NTS 73P-16 and 74A-1); *in* Summary of Investigations 1984, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 84-4, p21-31.

__________ (1985): Geology of an area around Star Lake; Sask. Energy Mines, Open File Rep. 85-6, 83p.

__________ (1987): Gold deposits of the La Ronge Domain; Geol. Assoc. Can., Field Trip Guide 4, Joint Annual Meeting with Mineral. Assoc. Can., Saskatoon, Sask., 70p

Field Observations on the Broken Hammer Gold Occurrence and NIKO Gold Deposit[1]

P.L. Schwann

Schwann, P.L. (1991): Field observations on the Broken Hammer gold occurrence and NIKO gold deposit; *in* Summary of Investigations 1991, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 91-4.

Studies of lithologic, structural and alteration characteristics of gold occurrences along the Byers Fault began in 1989, (Schwann, 1989) and continued in 1990 (Schwann, 1990) and 1991. Two prospects, Broken Hammer and NIKO, were investigated in 1991. They are located 32 km apart at opposite ends of the Byers Fault (Figure 1).

Previous work in the area by the Saskatchewan Geological Survey was carried out by Byers (1948), Kirkland (1959), Johnston (1969, 1970), and more recently by Harper (1984, 1985). Coombe (1984) presents an overview of gold deposits in the Star-Waddy Lakes area of the La Ronge Domain.

1. Regional Geology

The Byers Fault area is located within the La Ronge Domain near Waddy Lake, about 150 km northeast of La Ronge. The geology has been described by Harper (1984, 1985) as follows. The metavolcanic succession comprises ultramafic to intermediate flows which are overlain by felsic flows and pyroclastics. Intravolcanic metagreywackes, meta-argillites meta-arkoses and iron formation locally overlie the felsic metavolcanics. This metamorphosed volcanosedimentary assemblage is unconformably overlain by metaconglomerates. The supracrustal rocks are intruded by a variety of ultramafic to granitic plutons and stocks ranging in age from synvolcanic to post-tectonic. The dominant structural feature of the area is the 35 km long, easterly-trending Byers Fault. In the west, it dips moderately to steeply south but at its eastern end it dips steeply northwest. Subparallel faults (e.g. Oven-Corner Lakes Fault) form splays to the south of the main structure and intersect it in the Lower Waddy Lake area. Harper (1985) envisaged these faults as early ductile brittle structures formed parallel to intrusion margins during D1 folding. Northwest-trending lineaments offset the Byers Fault, which at its east end is deflected to the northeast by the north-northeast-striking Redhill Lake Tectonic Zone (RLTZ), the northern extension of the McLennan Lake

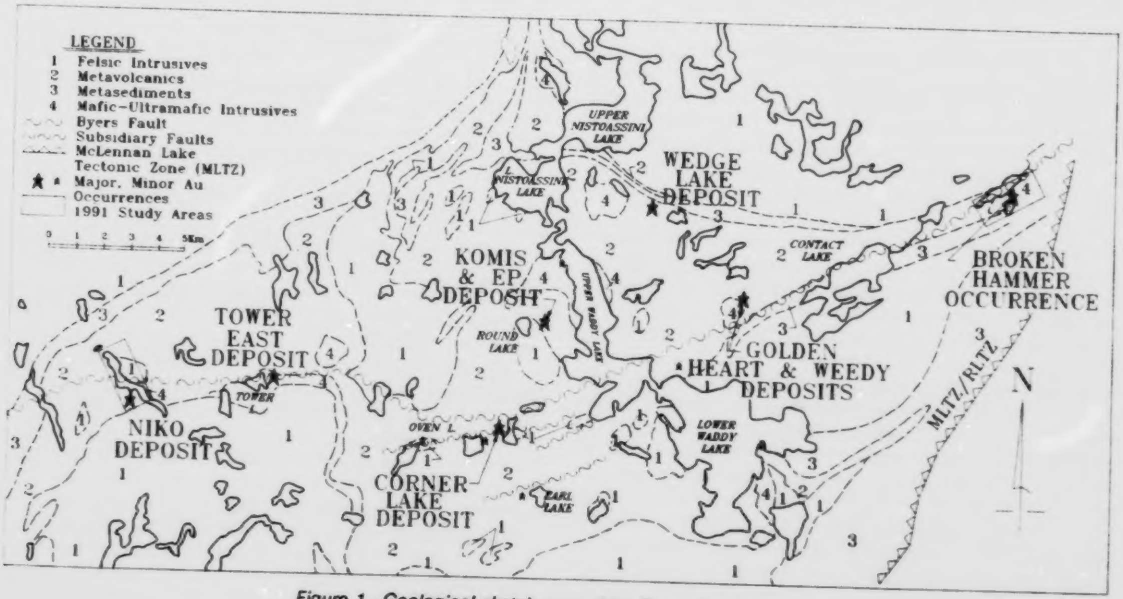

LEGEND
1 Felsic Intrusives
2 Metavolcanics
3 Metasediments
4 Mafic-Ultramafic Intrusives
Byers Fault
Subsidiary Faults
McLennan Lake Tectonic Zone (MLTZ)
Major, Minor Au Occurrences
1991 Study Areas
0 1 2 3 4 5Km
UPPER NISTOASSINI LAKE
NISTOASSINI LAKE
WEDGE LAKE DEPOSIT
CONTACT LAKE
BROKEN HAMMER OCCURRENCE
KOMIS & EP DEPOSIT
UPPER WADDY LAKE
ROUND LAKE
TOWER EAST DEPOSIT
TOWER
GOLDEN HEART & WEEDY DEPOSITS
NIKO DEPOSIT
OVEN L.
CORNER LAKE DEPOSIT
EARL LAKE
LOWER WADDY LAKE
MLTZ/RLTZ
N

Figure 1 - Geological sketch map of the Byers Fault area.

---

(1) Saskatchewan Project A.144, from which this report derives, was funded in 1991 under the Canada-Saskatchewan Partnership Agreement on Mineral Development 1990-95.

Tectonic Zone (MLTZ) (Harper, 1985). The MLTZ was interpreted by Lewry (1983) as a possible thrust zone, and by Poulsen *et al.*, (1987) as a transpressional shear, separating Central Metavolcanic Belt rocks from MacLean Lake Belt Gneisses.

Regionally the area shows geological similarities to gold-producing camps in the Superior Province (Colvine *et al.*, 1988). For example gold deposits are located:

1. within or immediately adjacent to volcanic belts intruded by felsic plutons,
2. within regional deformation zones,
3. close to faults and fault splays, and
4. within rocks metamorphosed at greenschist to middle amphibolite facies.

The Byers Fault zone was first recognized as a gold-bearing structure in the early sixties from work done by the Augustus Concession (SEM Assessment Files 64D4-NE-0001 and -0005). Numerous showings are located along the Byers and related faults (Figure 1). The majority are structurally controlled quartz-sulphide-carbonate-healed shear and tension veins within volcanics and felsic to mafic intrusions (Coombe *et al.*, 1985).

2. Broken Hammer Property

a) Property History

The Broken Hammer property is located 165 km northeast of La Ronge, 5 km west of Highway 102 and is centred on Alligator Lake. The area was initially explored in 1967 by Sherritt Gordon Mines Ltd. for its base metal potential. Six diamond drill holes tested a 2 mile long series of northeast-trending ground electromagnetic conductors. Disseminated to massive pyrite-pyrrhotite mineralization in metavolcanic rocks was intersected over a 0.5 m core length but no further drilling was done. In 1982, Saskatchewan Mining Development Corporation (SMDC) identified two strong and two weak electromagnetic conductors along strike immediately southwest of the property. A grab sample of a garnet-actinolite-chalcopyrite rock (interpreted as silicate facies iron formation) on strike with one of the strong conductors assayed 0.56 percent Cu and 310 ppb Au (Assessment File 64D4-0080). Alligator Resources acquired the area in January, 1985. Initial prospecting of the northern property (CBS 6120) resulted in an anomalous Pd assay (4000 ppb), which could not be repeated. Subsequent ground prospecting discovered several interesting showings, including a massive sulphide boulder (up to 3800 ppm Zn), an asbestos vein in an ultramafic body and an auriferous quartz vein (the "Broken Hammer" occurrence) which assayed up to 21 oz/ton Au. Recent exploration programs on the property have included prospecting, trenching, sampling (rock, biogeochemistry, soils) and minor outcrop stripping; however, the Broken Hammer gold occurrence remains the most promising discovery to date.

b) Geology of the Broken Hammer Property

The property area is underlain by Central Metavolcanic Belt rocks of the La Ronge Domain. Ultramafic to felsic metavolcanics intruded by ultramafic and mafic sills are structurally underlain by meta-arkose and calcareous metagreywackes. To the northeast the granitic Nistoassini Pluton intrudes the metavolcanic sequence and a diorite-gabbro body. A related granodioritic sill intrudes the metavolcanics to the north of Radiant Lake.

The contact between the metavolcanics and metasediments, although not exposed, is marked by a prominent topographic lineament suggesting it may be a faulted contact. Regionally, Harper (1985) envisaged the faulted contact between the metavolcanics and structurally underlying metasediments as the faulted limb of a syncline, thereby implying that the metasediments are younger than the metavolcanics. The Byers Fault itself does not outcrop, although its trace is indicated by a strong positive vertical gradiometer signature, passing through the long axis of Alligator Lake. Northeast and southwest of Alligator Lake it is marked by a muskeg-filled lineament. Gneisses of the MacLean Lake Belt outcrop directly to the east of the map area, separated from the Central Metavolcanic Belt sediments by a series of topographic ridges. The property was mapped at 1:8600 scale (Map 1), the showing itself at 1:50 scale.

Metavolcanics

Metavolcanics occur both as flows and pyroclastics and are compositionally variable from ultramafic to felsic. Intermediate to felsic metavolcanics are locally garnetiferous. Interlayered amphibole-garnet rocks may represent thin, intravolcanic silicate facies iron formations.

Ultramafic flows form extensive, smooth weathering, rounded outcrops that weather a distinctive emerald green. They are fine grained and massive, commonly crosscut by hairline-width, white-weathering veinlets.

Mafic metavolcanics are not abundant. To the south of Radiant Lake they form hills of gray, massive, hornblende porphyroblastic rock and east of Alligator Lake appear as dark gray to black, fine-grained, massive, hornblende-phyric flows interlayered with intermediate metavolcanics.

Intermediate metavolcanics occur east and west of Alligator Lake. They are grayish green and texturally variable, from fine-grained, laminated, aphyric flows and tuffs to amphibole and plagioclase-phyric, unsorted lapilli tuffs containing both mafic and felsic fragments. Intermediate tuffs are commonly interlaminated with thin, (<0.5 m) felsic layers. Felsic metavolcanics outcrop along the southwest shore of Alligator Lake where they are pink weathering, quartz phyric, massive units that host heavily disseminated (10 percent) pyrite mineralization. Sericite schists interlayered with intermediate lapilli tuffs outcrop along the west shore of Alligator Lake along strike of the felsic metavolcanics.

Metasediments

Metasediments, including meta-arkose and calcareous metagreywackes, outcrop as ridges in the eastern part of the study area, structurally underlying the metavolcanics and ultramafic intrusives.

Meta-arkoses comprise thick, buff to pink weathering, fine-grained, massive units characterized by centimetre-scale fracture cleavage. They contain minor muscovite and biotite and thin, (centimetre-wide) diopsidic bands. Massive, white, discontinuous, brittle quartz veins cut the meta-arkose. Both the quartz veins and wallrock contain disseminated (2 percent) pyrite. The meta-arkoses are structurally overlain by dark gray to green, laminated calcareous metagreywackes. Layers containing mafic fragments suggest a volcanic source for the metagreywackes which are composed of quartz, felspar, biotite, amphibole, epidote and finely disseminated sulphide.

Ultramafic Intrusives

Ultramafic rocks, including massive to weakly foliated peridotite, pyroxenite and tremolite schists, outcrop in two bodies. In both, relict oikocrystic clinopyroxene enclosing olivine, is preserved. The larger body of peridotite intrudes the metavolcanics between Alligator and Radiant Lakes, truncating them along strike. It is bluegreen with orange knobbly weathering, and medium to coarse grained. A centimetre-wide asbestos vein crosscuts the intrusive. The second, compositionally and texturally similar ultramafic body outcrops to the east of the main body. It is dominantly medium to coarse grained and shows distinct centimetre- to metre-scale compositional layering of peridotite (blue-green weathering) and pyroxenite (orange weathering). The second body is concordant to the metavolcanic sequence and structurally underlies the green-weathering ultramafic metavolcanics. Primary magmatic layering (S_0) is parallel to interlayering of the intermediate and felsic metavolcanics. A thin (2 m wide) zone of cream weathering, massive tremolite schist may represent an altered, thermal contact zone between the peridotite and overlying ultramafic metavolcanics. Magnetite, formed by serpentinization of olivine, occurs in peridotite and to a lesser extent in pyroxenite. Tremolite schists are non-magnetic. Sulphides and chromite were not observed in the ultramafic bodies. The position and close spatial association of the ultramafic intrusions with the ultramafic basalts suggests the intrusions are subvolcanic feeders, as proposed by Harper (1985).

Mafic, Intermediate & Felsic Intrusives

The large granitic intrusion that outcrops along the northwest shore of Alligator Lake is part of the Nistoassini Pluton. It is pinkish-white weathering, homogeneous, moderately foliated, medium grained, and contains quartz, potassium felspar and biotite as major constituent minerals. The intrusion is in sharp, folded contact with metavolcanics on the west shore of Alligator Lake where both are garnetiferous.

A smaller, granodioritic sill intrudes the metavolcanics midway between Alligator and Radiant Lakes. Quartz veining is focussed along the hanging wall contact of the granodiorite and metavolcanics. Both the granodiorite and quartz are strongly foliated to submylonitic and are sericitic and pyritic.

A gabbroic to dioritic stock occurs at the north end of Alligator Lake, adjacent to the Nistoassini Pluton. It is dark green to black weathering, coarse-medium grained, and locally displays compositional layering defined by centimetre-scale, white plagioclase- and dark green clinopyroxene-rich bands. Gabbro-diorite contacts are gradational and marked by a decrease in mafic content. The relationship between the gabbro-diorite body and Nistoassini Pluton is uncertain; however, Harper (1984) noted, that in places, the pluton has a gabbroic-dioritic marginal phase.

c) Alteration

Some rocks in the area have been subject to alteration prior to metamorphism. Hydrothermal alteration of the intermediate to felsic metavolcanic rocks is indicated by "mineralogical anomalies" (Allard and Carpenter, 1982). Local concentrations of garnet-sillimanite and sericite represent metamorphic manifestations of aluminous and potassic alteration respectively.

d) Structure and Metamorphism

Throughout the area, the dominant tectonic foliation (S_1) generally parallels primary layering (S_0) and dips moderately to steeply (70° to 80°) to the northwest. Early isoclinal folds (F_1) with boudinaged limbs plunge steeply (70°) to the northeast. A younger generation of open folds (F_2) was noted in the meta-arkosic and diopsidic metasedimentary rocks. Fold axes plunge moderately to steeply (55° to 70°) to the north. Other linear fabrics, including amphibole and garnet mineral lineations and streaking of felspar-amphibole aggregates, plunge moderately to steeply (55° to 65°) to the north-northeast.

Faulting is indicated by steep, northeast-trending cliffs parallel to S_0/S_1 along the shore of Alligator Lake. A zone of strong deformation is indicated by an east-northeast-trending hartschiefer banding (millimetre scale) in ultramafic rocks to the southwest of Radiant Lake.

Garnet and sillimanite porphyroblasts in sericitized felsic volcanics indicate middle to upper amphibolite facies conditions. Crystallographic orientation of amphibole and dimensional orientation of garnet porphyroblasts in S_1/S_0 contrast with a random orientation of sillimanite porphyroblasts suggesting peak deformation occurred during or after amphibole and garnet blastesis, but before peak metamorphism indicated by sillimanite growth.

e) Mineralization

The Broken Hammer gold showing is a structurally-controlled quartz vein, emplaced into a sequence of interlayered intermediate to felsic metavolcanics and amphibole-garnet layers. The quartz vein is parallel to the regional S_1/S_0 foliation, and dips steeply (70° to 80°) to the northwest. At the discovery outcrop the quartz vein is exposed over a 15 m strike length, and ranges from 5 cm to 1.5 m wide with an average width of 1.2 m. The quartz is sheared and glassy and contains trace molybdenite, pyrite and visible gold. A thin, concordant sliver (10 cm) of granodiorite intrudes the hanging wall. Contacts of the quartz vein with hanging wall and footwall rocks are sheared but sharp; the strain fabric rapidly decreases outward over 0.5 to 1.0 m from the mineralized quartz vein. A reddish brown colouration in the immediate footwall reflects the distribution of fine (< 1 mm), disseminated (< 2 percent) garnets and pyrite. Garnets first appear 5 m from the footwall, and increase in size (5 mm) and abundance (15 percent) towards the quartz vein; pyrite is also more heavily disseminated in the footwall proximal to the quartz vein. Pyrite and garnet do not occur in the hanging wall rocks. At the base of the outcrop, heavily disseminated to semi-massive pods of molybdenite occur at the footwall contact of the quartz vein and metavolcanics. The quartz vein continues along strike to the northeast for another 250 m, beyond the discovery outcrop; however, here the quartz vein lies at the contact of a granodioritic sill and overlying intermediate metavolcanics. The granodiorite has a strong planar fabric defined by quartz ribboning and alignment of felspar and biotite. Strong linear fabrics are defined by biotite-felspar aggregates and elongate quartz grains. Both planar and linear fabrics are consistent with the regional trends. Within the shear zone the granodiorite is sericitic, and both it and the quartz vein contain disseminated (< 2 percent) pyrite and molybdenite. The mineralized quartz vein is not exposed at surface southwest of the discovery outcrop. Assays from various grab samples are indicated in Table 1. Sample locations are indicated on Map 1.

f) Discussion

1) Gold mineralization on the Broken Hammer property is structurally controlled. The quartz vein and shear system is northeast-trending, subparallel to the regional S_1/S_0 foliation and Byers Fault. The quartz vein occurs at the contact of granodiorite and metavolcanics over most of its exposed length; however, preliminary assays indicate the highest values occur where the quartz vein is hosted by metavolcanics.

2) Disseminated molybdenite and pyrite are associated with gold mineralization.

3) Wall rocks immediately adjacent to the mineralized vein are noticeably altered. At the discovery outcrop garnet alteration is concentrated in the metavolcanics footwall. This alteration may be associated with gold mineralizing fluids. Alternatively, the compositional anisotropy between pre-existing garnetiferous and nongarnetiferous units may have constrained deformation and focussed later migrating, gold-bearing fluid.

4) Metasediments on the east of the property contain other environments for possible gold mineralization, and include:
i)brittle, fracture-controlled, quartz veins and pyritized wallrock in the meta-arkose (southeast of Radiant Lake) and,
ii)stratabound, sulphidic, calcareous metagreywacke horizons (east of Radiant Lake).

5) Ultramafics offer limited potential for PGE or nickel exploration due to the lack of sulphides and primary oxides.

6) Intermediate and felsic metavolcanics on the southeast and west shores of Alligator Lake may have some potential for base metals, especially as there are several showings along strike to the northeast (Fleming Lake) and southwest (Contact Lake). Garnet, sillimanite and sericite together with heavily disseminated pyrite is suggestive of hydrothermal alteration system underlying volcanogenic massive sulphide deposits.

Table 1 - Geochemical Assays - Broken Hammer Property

Sample	Mo ppm	Cu ppm	Pb ppm	Zn ppm	Ag oz/t	Ni ppm	Co ppm	Mn ppm	Fe %	As ppm	Au oz/t
180	30	600	100	3200	.05	400	100	600	30.04	100	.004
181	10	100	100	100	.01	100	100	400	2.76	100	.002
177A	1397	162	75	10	.06	29	25	101	4.45	2	.002
177B	491	31	25	1	.03	9	3	44	1.09	2	.404
175	15790	300	100	100	.22	100	100	100	4.99	100	.023
178											.070
179	293	14	111	7	.10	33	5	204	1.22	14	.002

Samples 177A, 177B, 175 and 179 are from the Broken Hammer showing and strike extension; #178 from a quartz vein on the northwest side of Alligator Lake; #180 from a mineralized boulder; and sample 181 from mineralized quartz phyric rhyolite.

3. Kaslo/NIKO Property

a) Property History

The Kaslo property is located 130 km northeast of La Ronge. Exploration began in the late 1940s when E. Partridge found gold-bearing float. In the 1960s, Augustus Exploration Ltd. and Triana Exploration Ltd. discovered several gold occurrences along an easterly-trending fault zone at the contact between diorite and metavolcanics. A total of 6274 ft. was drilled with the better assays running 0.55 oz/ton over 5 ft. Since 1982, Golden Rule Resources, (operator) along with joint venture partners Goldsil Resources Ltd. and Cameco have undertaken exploration programs to define these gold occurrences. These have included geophysical surveys, mapping, prospecting and drilling programs. The NIKO occurrence was discovered in 1987 when a grab sample from a mapping program assayed 4620 ppb Au. Since then, exploration has concentrated on defining the potential of this occurrence.

b) Geology of the Kaslo Property

The Kaslo property is underlain by rocks of the Central Metavolcanic Belt of the La Ronge Domain. Interbedded mafic to felsic metavolcanics and sulphidic meta-argillites are intruded by mafic to felsic phases of the Brindson Lake Pluton and associated dykes. The dominant structural features of the area are reflected by topographic lineaments. From Tower Lake through Narrow Lake the Byers Fault is characterized by topographic depressions. Geophysically it coincides with a discontinuous zone of low magnetic signature and a broad (150 to 300 m wide) weakly conductive zone (Patterson, 1987). The zone of low magnetic signature bifurcates into east-northeast- and northeast-trending subzones east of the property. The dominant east-northeast-trending subzone transects the northern narrow neck of Narrow Lake. Its trace continues along the valley at the southwestern end of Narrow Lake and along the southern shore of Centre Lake. The southwest-trending subzone cuts across the south end of Narrow Lake. This early set of east-northeasterly- to north-easterly-trending magnetic lineaments is offset by a second set of northeasterly-trending lineaments and both sets by northwesterly-trending lineaments.

Exposure on the property is generally poor, characterized by small, blocky, isolated outcrops. The property was mapped at 1:2500 scale and the NIKO showing at 1:100 scale.

Metavolcanics

Metavolcanic rocks occur both as flows and pyroclastics and are compositionally variable from mafic to felsic. A mafic to intermediate unit outcrops along the south shore of Narrow Lake and is locally garnetiferous. It is dark grey to black weathering, aphanitic and well foliated with irregularly pitted surfaces making identification of primary features difficult. Quartz amygdales are locally abundant.

Intermediate pyroclastic rocks are relatively abundant in the west where they are interfingered with felsic metavolcanics. Lapilli tuffs are unsorted, massive and contain clast supported felsic and mafic lithic fragments. The dark green matrix is felspar-and amphibole-phyric. Both felsic fragments and the matrix contain 0.5 to 1.0 mm quartz eyes. These rocks become finer grained and matrix supported to the east. Clasts are elongated parallel to the foliation. A friable bluish-gray weathering, aphyric, vesicular, (scoriaceous ?) rock is poorly exposed at the southwest end of Narrow Lake. It shows a weak centimetre-scale fracture cleavage and contains fine biotite, quartz, felspar and graphite.

Felsic flows, along the south shore of Narrow Lake, form hills of massive, bluish-gray to pink weathering plagioclase- and amphibole-phyric rock that contains irregularly distributed quartz eyes and magnetite. Felsic lapilli tuffs generally occur interlaminated with intermediate pyroclastic rocks. They are plagioclase-phyric, weather cream to pale green with pitted surfaces and contain unsorted felsic and mafic fragments. Felsic metavolcanics are weakly to moderately foliated and clasts are strongly elongate. At the southwest end of Narrow Lake they host heavily disseminated pyrite mineralization.

Metasediments

Recessively weathered layers of sulphidic meta-argillite are interdigitated with metavolcanics in the central part of the map area. Meta-argillites are dark grey to rusty brown (gossanous), aphanitic, siliceous, finely laminated and moderately foliated. They contain finely disseminated pyrrhotite, pyrite and magnetite.

Mafic-Felsic Intrusions

Granodioritic to gabbroic intrusions occur on both the north and southeast shores of Narrow Lake. They are separated lithologically by a westward thickening wedge of supracrustal rocks and structurally by both the east-northeast-trending Byers Fault and a northwest-trending lineament that runs through Narrow Lake. The larger southern intrusive body is part of the compositionally zoned Brindson Lake Pluton; the northern body is compositionally and texturally similar and may represent a faulted off portion. Cream to buff weathering, massive, equigranular granodiorite is the most abundant phase of the Brindson Lake Pluton. Diorite, which varies from leucocratic to melanocratic, is a marginal phase outcropping along the southeast and northeast shores of Narrow Lake. It is a cream to medium gray, medium- to coarse-grained, homogeneous rock comprising plagioclase, clinopyroxene and amphibole.

The intrusive rocks are generally weakly foliated except adjacent to the narrows in Narrow Lake where discrete narrow shears in the granodiorite develop into a more extensive zone of shearing and alteration. Numerous dykes intrude the metavolcanics and granitic to gabbroic intrusives. A pale green weathering, massive, fine-grained, amphibole-phyric mafic dyke crosscuts the metavolcanics. Salmon-pink-weathering, aphyric to quartz-phyric; white-weathering plagioclase-phyric; and white-weathering, equigranular tonalite dykes intrude the supracrustal sequence and the marginal mafic phase of the Brindson Lake Pluton. The salmon-pink-weathering felsic dykes also intrude the granodiorite and may be related to a more felsic phase of the pluton.

c) Structure

Regional foliation parallels the trace of the Byers Fault and dips steeply (70° to 80°) to the south-southeast.

Early isoclinal minor folds were not observed, although the large scale distribution of lithologies and the convergence of foliation are suggestive of a major closure. Clast elongation in intermediate-felsic metavolcanics plunges moderately (50° to 60°) southeast. In drill core, the Byers Fault is a 1 to 15 m wide zone of deformation and alteration within granodiorite to the north of Narrow Lake. It is characterized by a quartz-carbonate vein stockwork and a chloritic fault gouge containing angular, altered, and foliated granodiorite clasts. Hematite and epidote coat fracture surfaces. These features also characterize the fault zone where it outcrops on the north shore of Narrow Lake at the narrows. Here the granodiorite is strongly foliated and hematized. Quartz veining and silicification are well developed in the mafic to intermediate volcanics in the hanging wall of the fault.

The contact of the metavolcanics and granodiorite in the hanging wall of the Byers Fault to the north of Narrow Lake is a zone of brecciation known locally as the "Fault Breccia". It is not exposed at surface. A number of west-northwest-trending shears and faults occur along the southern shore and are crosscut by northwest-trending narrow, centimetre-wide shears.

d) Mineralization

Drilling has identified three zones of mineralization: the "Cornflake", "Red Cube" and "Granny" at the NIKO occurrence (Golden Rule Resources Annual Report, 1990). All are in the hanging wall of the Byers Fault. Relative to the "Fault Breccia", the "Cornflake" and "Red Cube" zones occur in the hanging wall of the "Fault Breccia" and the "Granny" zone in the granodiorite footwall.

Cornflake Zone

The "Cornflake" zone is up to 25 m wide and contains most of the identified reserves of the NIKO occurrence. Mineralization is hosted by quartz-amphibole-pyrite veins (< 10 cm width) and enclosing silicified and pyritized, strongly foliated mafic to intermediate metavolcanics. Uniformly disseminated (10 percent), 0.5 to 1.0 mm pyrite grains flattened in the foliation plane account for its distinctive "cornflake" texture. Gold grade correlates sympathetically with pyrite content.

Red Cube Zone

"Red Cube" mineralization is restricted to centimetre-wide quartz veins that are oblique to the regional foliation. Mineralized zones are up to 2 m wide. Gold occurs as individual, isolated grains or as fine disseminations peppered throughout brick-red, hematized pyrite cubes in the quartz veins.

Granny Zone

The "Granny" Zone is hosted by bleached albitized granodiorite. Mineralized zones are narrow, and free gold occurs in silicates and along discrete hairline fractures. Sulphides or their oxidized equivalents are absent.

All three styles of mineralization have been tentatively identified at surface by the author. "Cornflake" and "Red Cube" mineralization occurs at the exposed NIKO showing. A 0.5 to 1.0 m wide, east-northeast-trending, steeply south-dipping, quartz-infilled shear zone separates pyritized felsic metavolcanics (herein referred to as the "Frosted Flake" style) from pyritized mafic to intermediate metavolcanics ("Cornflake" style) over a 65 m strike length (Figure 2). The quartz vein contains green amphibole rosettes (<20 percent), and fine (<0.5 mm) to coarse (3 cm) pyrite grains and clusters. The vein is locally brecciated along its footwall contact. A north-northwest-trending, narrow, (<0.5 m wide) zone of brick-red hematite ("Red Cube" style) crosscuts the quartz vein and fault breccia. Mineralization similar to that of the "Granny" zone is exposed on an island of bleached granodiorite just west of the narrows of Narrow Lake. The outcrop is transected by hairline fractures, epidote-filled shears and a quartz-phyric felsic dyke, but lacks sulphide mineralization and/or quartz veining. Visible gold was found enclosed within a biotite crystal in the granodiorite. Significantly, this outcrop is located in the footwall of the Byers Fault.

"Cornflake" and "Frosted Flake"-style mineralization is exposed at the southwest shore of Narrow Lake at the PW showing. A metre-wide shear zone in pyritized felsic metavolcanics comprises powdery-white quartz and albite alteration containing heavily disseminated to semi-massive pyrite mineralization. A 30 to 40 cm quartz vein is central to the shear zone. The zone continues along strike to the west for at least 100 m where it is exposed in three old, overgrown trenches. Intermediate metavolcanics forming the footwall to the sheared, felsic metavolcanics exhibit flattened disseminated (5 to 10 percent) pyrite similar to that in the "Cornflake Zone" proper. Assay results from the NIKO surface showing and the PW showing are listed in Table 2.

Table 2: Gold Assays - NIKO showing

Sample #	oz/t
120	.001
154	.009
157	.004
158	.001
D9	.047
D22A	.195
D22B	.014
DFB	.352

D sample locations are indicated on Figure 2 and on the accompanying map.

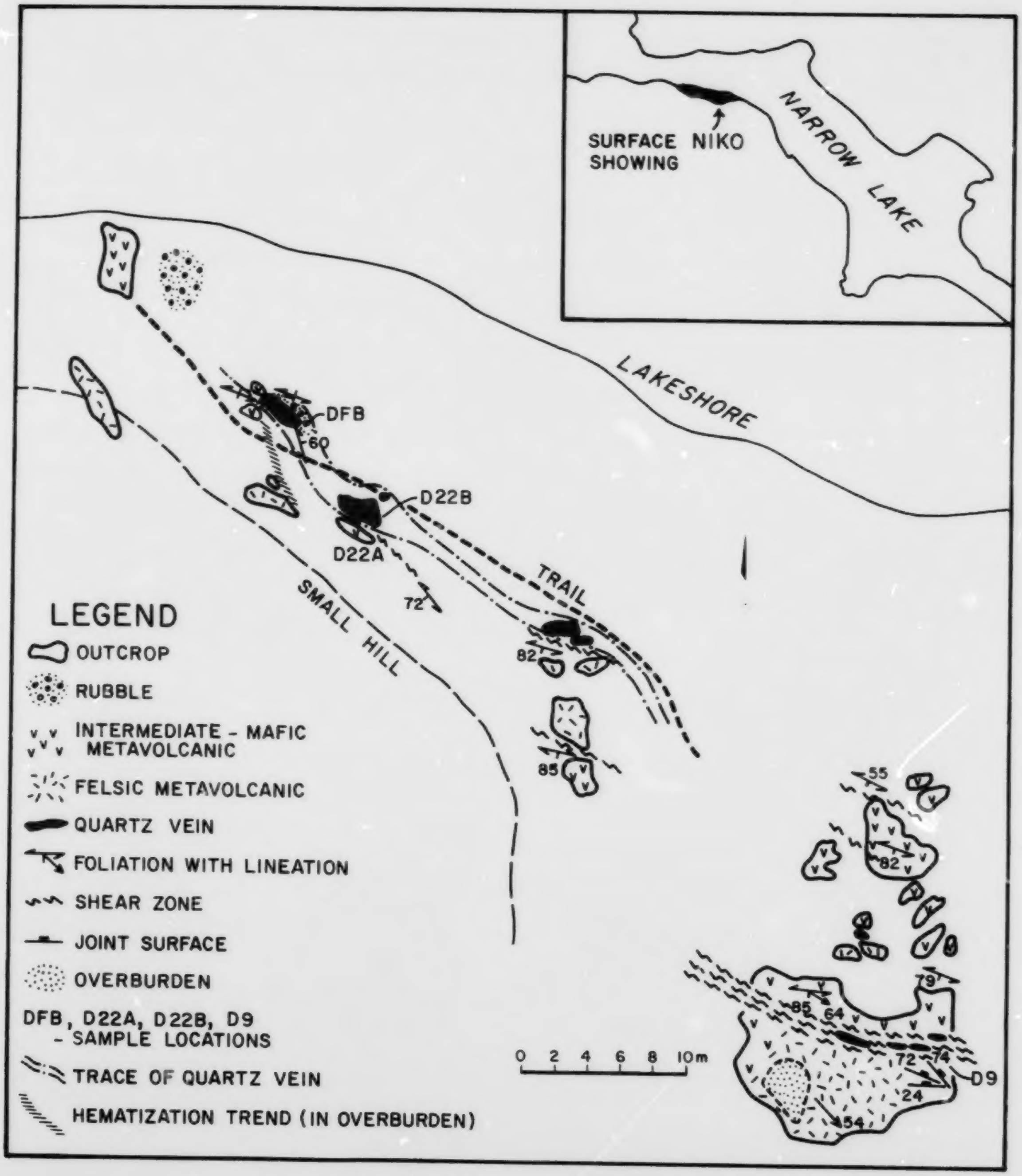

SURFACE NIKO SHOWING
NARROW LAKE
LAKESHORE
TRAIL
SMALL HILL
DFB
60
D22B
D22A
72
82
85
55
82
79
85
64
72
74
D9
24
54
0 2 4 6 8 10m
LEGEND
OUTCROP
RUBBLE
INTERMEDIATE - MAFIC METAVOLCANIC
FELSIC METAVOLCANIC
QUARTZ VEIN
FOLIATION WITH LINEATION
SHEAR ZONE
JOINT SURFACE
OVERBURDEN
DFB, D22A, D22B, D9 - SAMPLE LOCATIONS
TRACE OF QUARTZ VEIN
HEMATIZATION TREND (IN OVERBURDEN)

Figure 2 - Geological sketch map of the NIKO showing.

e) Discussion

1) Gold mineralization on the Kaslo property is controlled by a combination of structure and alteration. The "Cornflake", "Red Cube" and "Granny" zones have different structural and alteration characteristics. "Cornflake" style mineralization is early, with pyrite and quartz veins confined to zones which parallel the regional east-northeast-trending foliation and the Byers Fault. The sympathetic correlation between gold and pyrite suggests that gold mineralization was associated with sulphidation and the pyrite grains flattened within the regional foliation suggest this process was pre- to syn-tectonic. "Red Cube"-

style mineralized structures trend northwest and crosscut "Cornflake" mineralization. A later episode of hematization has oxidized primary pyrite. The nature and timing of the "Granny" mineralization is difficult to assess as there is no obvious structural or alteration control. From this summer's work it was observed that "Granny"-style mineralization is not confined to the hanging wall of the Byers Fault.

2) A second example of the "Cornflake" and "Frosted Flake" style mineralizations occurs at the PW showing at the southwest corner of Narrow Lake. Sample 158 is from the altered felsic rocks within the shear zone. The low assay result indicates that at this location the shear zone is not auriferous. Assays for "Cornflake"-style mineralization footwall to the shear zone and from trenches along strike of the shear zone were not available at the time of publication.

3) Other auriferous environments include;
(i) stratabound sulphidic meta-argillite units, and
(ii) felsic and mafic phases of the Brindson Lake Pluton that are crosscut by structures. The latter typify the mineralization style of the original Kaslo occurrence.

4) Mylonitized fragments in the Byers Fault breccia indicate early ductile movement overprinted by an episode of brittle deformation. Offset of the Byers Fault by later northwest-trending structures is indicated on surface and by geophysical lineament patterns.

4. Conclusions

The Byers Fault structure trends east-northeast at the Kaslo property and northeast at the Broken Hammer Property. The trend of this regional structure is paralleled by mineralized, lower order structures at these locations. Compositional anisotropies are favourable sites for shear zone and quartz vein development. Mineralized shear structures have brittle-ductile transitional characteristics. Altered wallrock indicates fluids penetrated and interacted with rocks outside the shears. Gold mineralization is commonly found associated with sulphides (and their oxidized equivalents), including pyrite and molybdenite, an exception being "Granny"-style mineralization, where sulphides are notably absent.

5. Acknowledgments

Company support continued to complement this project. Radiant Club members at Alligator Lake enthusiastically extended gracious hospitality; Golden Rule Resources Ltd. continued to provide logistical support. Wayne Laturnas provided mapping assistance. Discussions with provincial colleagues T. Sibbald, D. Thomas, B. Slimmon and federal colleagues H. Poulsen and E. Froese provided additional insight into environments of gold and base metal mineralization.

6. References

Allard, G.O. and Carpenter, R.H. (1982): The relationship between aluminosilicate assemblages and sulphide mineralization in volcanic terranes; 1982 - Geol. Assoc. Can., Prog. Abstr., v7, p35.

Byers, A.R. (1948): The geology of the Waddy Lakes area; Sask. Dep. Miner. Resour., Prelim. Rep. 1, 36p.

Colvine, A.C., Fyon, J.A., Heather, K.B., Marmont, S., Smith, P.M. and Troop, D.G. (1988): Archean gold deposits in Ontario; Ont. Geol. Surv., Misc. Pap. 139, 136p.

Coombe, W. (1984): Gold in Saskatchewan; Sask. Energy Mines, Open File Rep. 84-1, 134p.

Coombe, W., Lewry, J.F., and Macdonald, R. (1985): Regional geological setting of Gold in the La Ronge Domain, Saskatchewan; *in* Clark, L.A. (ed.), Gold in the Western Shield, CIM Spec. v39, p25-56.

Golden Rule Resources Ltd. (1990): Annual Report, 36p.

Harper, C.T. (1984): Geological mapping, Waddy Lake area (part of NTS 64D-4 and -5): *in* Summary of Investigations 1984, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 84-4, p6-20.

__________ (1985): Bedrock geological mapping, Waddy-Tower Lakes area (part of NTS 64D-4 and -5 and 74A-1 and -8); *in* Summary of Investigations 1985, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 85-4, p6-17.

Johnston, W.G.Q. (1969): The geology of the eastern portion of the Waddy Lake area, Saskatchewan; Sask. Dep. Miner. Resour., Rep. 127, 43p.

__________ (1970): The geology of the May Lake area (east half), Saskatchewan; Sask. Dep. Miner. Resour., Rep. 130, 36p.

Kirkland, S.J.T. (1959): The geology of the Brabant Lake area, Saskatchewan; Sask. Dep. Miner. Resour., Rep. 33, 31p.

Lewry, J.F. (1983): Character and structural relations of the "McLennan Group" meta-arkoses, McLennan-Jaysmith Lakes area; *in* Summary of Investigations, 1983, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 83-4, p49-55.

Patterson, W. (1987): Report on 1987 Geophysical Surveys, Kaslo Project, CBS 6417 & ML 5476 - NTS 74A-1; Assessment File 74A-1-0133

Poulsen, K.H., Ames, D.E., Galley, A.G., Derome, I. and Brommecker, R. (1987): Structural studies in the northern part of the La Ronge Domain; *in* Summary of Investigations 1987, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 87-4, p107-115.

Schwann, P.L. (1989): La Ronge gold belt geology: gold deposits in the Byers Fault area (part of 64D-4); *in* Summary of Investigations 1989, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 89-4, p41-43.

__________ (1990): Field Observations on the Corner Lake and Tower East gold showings, Byers Fault area; *in* Summary of Investigations 1990, Saskatchewan Geological Survey, Misc Rep. 90-4, p60-63.

Structural Controls and Geochemical Character of the Seabee Gold Mine, Laonil Lake, Glennie Domain[1]

D.J. Schultz[2] and R. Kerrich[2]

Schultz, D.J. and Kerrich, R. (1991): Structural controls and geochemical character of the Seabee Gold Mine, Laonil Lake, Glennie Domain; *in* Summary of Investigations 1991, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 91-4.

The Seabee gold deposit, located in the Glennie Domain, is one of a number of mesothermal precious metal deposits in northern Saskatchewan. This report summarizes previous geological work (Coombe, 1984; Delaney, 1991) as well as the preliminary results of research towards an M.Sc. degree by the senior author.

Initial objectives of the study were to conduct structural and lithological mapping to establish the geological setting of the Seabee deposit. Electron microprobe analyses of minerals and whole rock geochemistry will characterize the rocks and their alteration. Fluid inclusion, stable isotope and radiogenic isotope studies will help determine the source and composition of the hydrothermal fluids and solutes, and to constrain the method and timing of gold precipitation. Collectively, these techniques will be used to assess the Pressure-Temperature-time-fluid (P-T-t-f) history of host-rock lithologies and mineralization, and determine the deposit's origin.

1. Regional Geology

The Proterozoic Trans-Hudson Orogen comprises a series of lithotectonic domains including the Flin Flon, La Ronge and Glennie domains (Lewry and Sibbald, 1977). Lewry (1981) suggested that the Glennie Domain was a remnant Archean micro-continent that collided with the La Ronge volcanic arc. U-Pb age determinations have revealed that the only Archean rocks found so far are in the Iskwatikn Dome, near the western margin of the Glennie Domain (Bickford and Van Schmus, 1985; Van Schmus and Bickford, 1984; Van Schmus *et al.*, 1987; Chiarenzelli, 1989; Delaney *et al.*, 1988, 1990). As further detailed structural studies have been completed, the northern part of the Domain has been interpreted to be a stack of southeastward directed, refolded allochthonous thrust sheets formed during terminal collision in the Trans-Hudson Orogen (Bickford *et al.*, 1990; Lewry and Collerson, 1991).

The Seabee deposit is hosted by mylonitic shear zones which transect the Laonil Lake Intrusive Complex (LLIC) (Delaney, 1986). The LLIC, outcropping over a strike length of 4 km, comprises dominantly gabbro and diorite but contains late intrusive phases such as granodiorite, feldspar porphyry, mafic dykes ("green dyke") and intermediate dykes ("grey dykes") (The terms in parenthesis are field descriptions and in all cases, for any given rock type, the prefix "meta"- is assumed). Remnants and xenoliths of volcanic sedimentary and ultramafic rocks are also present. Emplacement of the LLIC is thought to post-date the D_1 and pre-date D_2 deformational events, which include development of the Ray Lake Synform and the emplacement of the post D_2 Eyahpaise Lake Pluton (see Figure 1 and Table 1) (Lewry, 1977). A sample of diorite collected by Delaney has been dated by U-Pb zircon at 1889 ±8.7 Ma (Chiarenzelli, 1989). To the northeast, the LLIC is unconformably overlain by Pine Lake volcanics from which a sodic rhyolite has been dated at 1832 ±2 Ma (Delaney *et al.*, 1990). To the west the LLIC is bounded by apparently earlier granodioritic to dioritic gneiss. The younger Eyahpaise (1859 ±5 Ma; Bickford *et al.*, 1986) tonalitic pluton lies to the south of the LLIC.

The Laonil Lake Shear Zone (LLSZ) (Delaney, 1986), a large scale regional lineament of intense deformation (D_2 or post D_2) characterized by protomylonites, occurs between the Eyahpaise Pluton and the LLIC (Figure 2). The D_3 deformational event, characterized by open folding in the Glennie Domain, overprints the entire region (Lewry, 1977).

2. Shear Structures

A network of anastomosing, sub-parallel shear structures (Seabee Shear Structures or SSS) crosscut the LLIC at an orientation of 045° to 085°, dipping sub-vertically or steeply northerly (Figure 2). Three discrete sets of structures have been recognized, which are oriented at 070°, 085° and 045° respectively. The 070° structures contain ore (Zone 2B, Zone 5 and Zone 161), whereas the 045° and 085° structures contain minor mineralization or are barren (Zone 20, Zone 18 and Zone 15). Feldspar porphyries occur in close proximity to many of the major structures (Zone 2 and Zone 20) as do contacts between gabbros and diorites and ruptured dykes (Zone 2, Zone 20, Zone 15 and east boundary Zone; Figure 2) suggesting that lithologic heterogeneities focussed deformation. The SSS and LLSZ share common orientations and similar kinematic indicators; consequently the SSS may represent a second order structure related to the LLSZ.

---

(1) Saskatchewan Project A.145, from which this report derives, was funded in 1991 under the Canada-Saskatchewan Partnership Agreement on Mineral Development 1990-95.

(2) Department of Geological Sciences, University of Saskatchewan, Saskatoon, Saskatchewan, S7N 0W0.

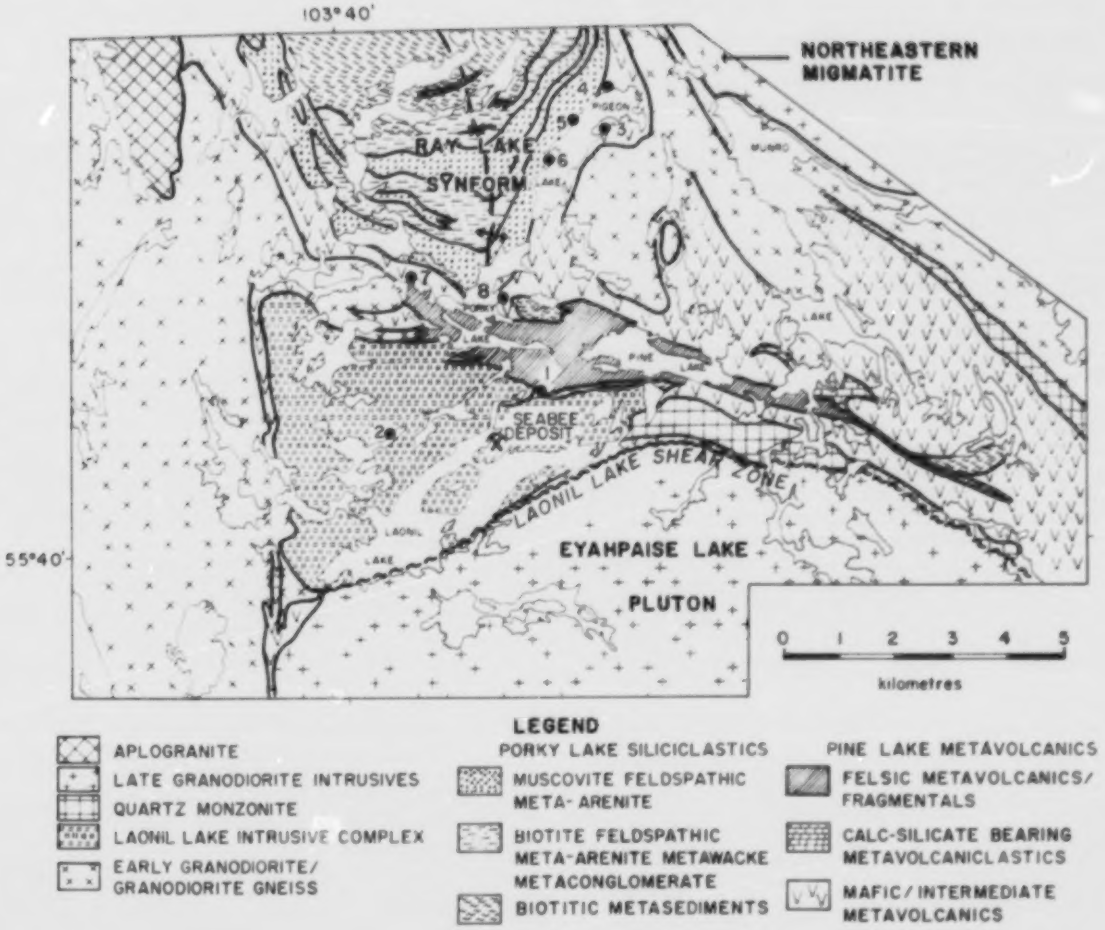

103° 40'
55°40'
NORTHEASTERN MIGMATITE
RAY LAKE SYNFORM
PIGEON
MUNRO
PORKY LAKE
PINE LAKE
SEABEE DEPOSIT
LAONIL LAKE SHEAR ZONE
LAONIL LAKE
EYAHPAISE LAKE PLUTON
1
2
3
4
5
6
7
8
0 1 2 3 4 5
kilometres
LEGEND
APLOGRANITE
LATE GRANODIORITE INTRUSIVES
QUARTZ MONZONITE
LAONIL LAKE INTRUSIVE COMPLEX
EARLY GRANODIORITE/ GRANODIORITE GNEISS
PORKY LAKE SILICICLASTICS
MUSCOVITE FELDSPATHIC META-ARENITE
BIOTITE FELDSPATHIC META-ARENITE METAWACKE METACONGLOMERATE
BIOTITIC METASEDIMENTS
PINE LAKE METAVOLCANICS
FELSIC METAVOLCANICS/ FRAGMENTALS
CALC-SILICATE BEARING METAVOLCANICLASTICS
MAFIC/INTERMEDIATE METAVOLCANICS

Figure 1 - Geological sketch map of the Laonil Lake map area illustrating the distribution of lithologies, shear structures, major Au showings and location of the Seabee gold deposit (after Delaney, 1986).

Table 1 - Correlation of Regional Geological and Local Seabee Events.

	Regional Events*		Seabee
D_1	Early folding event	D_1	Remnant Basalt
D_2	Laonil Lake Intrusive Complex ? Ray Lake Synform Eyahpaise Pluton Laonil Lake Shear Strucutre	D_2	Mafic dykes Felsic dykes V_1 stage Seabee Shear Structure Feldspar Porphyry V_2 stage Mineralization and ppt. Au
D_3	Wapassin Fold System	D_3	Re-activation of Seabee Shear Srocture (?) Brittle deformation of vein system Late Fractures (Aquifers) Mineralization and ppt. of sulphides Remobilization of Au (?)
Recent	Late Aquifer		V_3 stage (Barite and Calcite+Clay alteration) Remobilization of Au (?)

*Based on Lewry (1977)

According to Helmstaedt (1986, 1987), the LLSZ and SSS were dominated by coaxial deformation, or pure shear: this interpretation stems from the absence of any clear kinematic indicators of simple shear. Structural studies in other areas (Lister and Snokes, 1984) have demonstrated that mylonites formed in proximity to specific boundaries, such as a pluton margin, will be dominated by coaxial deformation. Thus, the SSS may have a dominant pure shear component, because the LLSZ and the Eyahpaise Lake Pluton conform with these criteria.

Subsequent field observations (Schultz, 1990) determined that disrupted gabbros or mafic flows in the vicinity of Zone 20, ex-

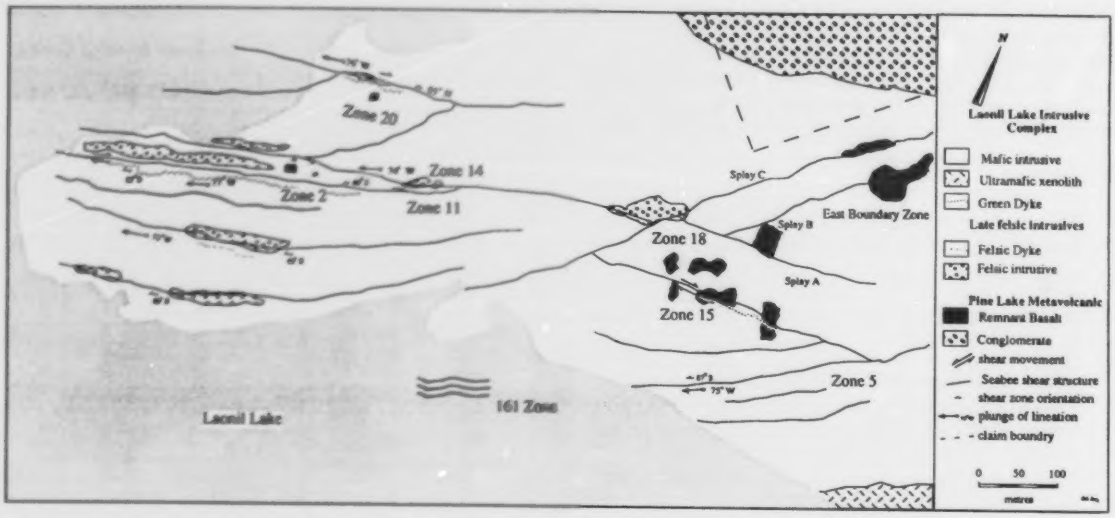

N
Laonil Lake Intrusive Complex
Mafic intrusive
Ultramafic xenolith
Green Dyke
Late felsic intrusives
Felsic Dyke
Felsic intrusive
Pine Lake Metavolcanic
Remnant Basalt
Conglomerate
shear movement
Seabee shear structure
shear zone orientation
plunge of lineation
claim boundry
0 50 100
metres
Zone 20
Zone 14
Zone 2
Zone 11
Splay C
Splay B
Splay A
East Boundary Zone
Zone 18
Zone 15
Zone 5
161 Zone
Laonil Lake
75° W

Figure 2 - Simplified geological map of the Seabee property illustrating the orientation of the Seabee Structure and major lithologies.

perienced non-coaxial deformation. Adjacent to zone 20, fragments show progressive flattening and rotation passing into the deformation zone in accordance with a simple shear model (Figure 3). In addition, weakly asymmetrical augen feldspar crystals were observed in the feldspar porphyry unit associated with Zone 2 (c.f. Bursnall, 1989). The augen structures are best developed in the vertical plane, although the asymmetry is not sufficiently distinct to determine sense of movement. A well developed mineral stretching lineation defined by the preferred alignment of hornblende and biotite, plunging at 80°W, implies a dip slip component to the SSS.

Lewry (1977) suggested that during D_2 deformation, the LLIC was trapped between the fold core of the Ray Lake Synform and the rising Eyahpaise diapir, thus acting as a resistant mass similar to a "mega-porphroblast". In this strain regime, the LLIC experienced both flattening (pure shear) and non coaxial deformation concentrated along the margin and confined internally within specific zones of weakness. Thus, in addition to pure shear, simple shear may have contributed to the deformation history of the LLIC.

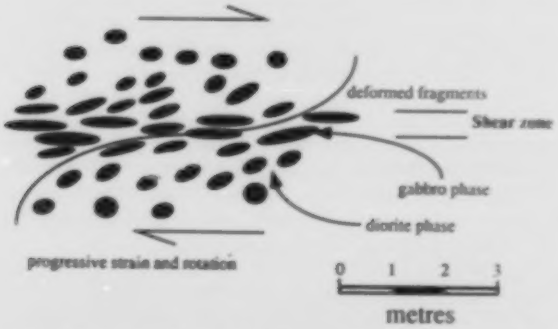

deformed fragments
Shear zone
gabbro phase
diorite phase
progressive strain and rotation
0 1 2 3
metres

Figure 3 - Ruptured unit within Seabee Shear Structure showing progressive strain and rotation of fragments (i.e. no-coaxial deformation).

3. Shear Zone Alteration

Progressive changes in both grain size and mineralogy occur along traverses from the least-deformed gabbro/diorite country rocks to their deformed and metasomatized counterparts in the shear zones. The following descriptions characterize these changes in four stages:

1. Outside the shear zone: massive gabbro/diorite is (grain size >500 μm) mechanically isotropic, relatively undeformed, with little or no biotitic alteration.

2. Margin of shear zone: gabbro/diorite with a markedly decreased grain size (300 to 200 μm) becoming foliated with a corresponding increase in modal biotite content.

3. Within the shear zone: biotite-actinolite-chlorite-carbonate alteration assemblage with a moderately to intensely developed foliation.

4. Shear zone core: segregated mafic (hornblende-actinolite-chlorite-biotite) and felsic (quartz-muscovite-biotite-feldspar) alteration assemblages (Cambell, 1986) with mylonitic fabric.

4. Vein Geometry and Petrography

Generally, quartz veins are only prominently developed within or adjacent to the shear zones. External to the shears, individual barren veins are sparse and of limited extent. In addition to early veins (V1), four sets of veins

(V2a-V2d) are associated with the SSS. These vein sets developed contemporaneously with shearing and contain gold where there are significant concentrations of sulphides. Late flat faults, crosscut the orebody (Zone 2). Clay minerals and barite are commonly precipitated as late fracture fillings (V3) associated with these faults.

First Stage:

V1 First stage veins, with no preferred orientation, predate shearing, and consist dominantly of quartz and albite.

Second Stage, V2:

a) Shear veins (c.f. Poulsen, 1989)
Veins are parallel to the shear foliation and are boudinaged in both the vertical and horizontal plane. Early tourmaline is developed parallel to the foliation. Later tourmaline and disseminated actinolite, chlorite, and pyrite are concentrated at or near the vein/wall rock boundary.

b) Extensional veins (c.f. Poulsen, 1989)
Veins are sub-parallel (60° to 80°) to shear foliation and typically folded about axial planes striking 065° to 085°. These veins contain a similar mineral assemblage as V2a and with progressive strain and rotation may become shear veins (V2a).

c) Extensional veins external but proximal to the shear zone (c.f. Poulsen, 1989)
Extensional veins, crosscut shear foliation at 90° to 120° and contain only minor tourmaline with limited sulphide (i.e. pyrite).

d) Breccia Vein (seen at Zone 14)
Host rock in proximity to breccia veins, are pervasively silicified with intense alteration represented by K-feldspar. The veins contain quartz, chlorite, muscovite, and albite with minor amounts of tourmaline and epidote. Host rock at or near vein boundaries contain actinolite, chlorite, and muscovite, with minor biotite and pyrite.

Late Stage

V3 Late Fractures
Late flat-lying fractures perpendicular to the shear structure (Figure 4), crosscut the orebody, and cause major water problems underground (where intersected by underground stopes water pressure was recorded at 100 to 200 psi and a flux of 200 to 300 gallons/day). In some places, clay minerals (kaolinite, illite and montmorilinite) and barite have precipitated in these late fractures. Where fluid has come into contact with vuggy pyrite and quartz (100 stope), calcite rhombs and free gold have precipitated in the open spaces.

a) Ore Petrography

Polished thin section S109-664.5, provided by Placer Development Limited, is a representative auriferous quartz vein sample from a diamond drill hole intersection with significant gold content (2.14 oz/ton). The polished section contains quartz-tourmaline- pyrrhotite-pyrite-chalcopyrite-bismuth-telluride and free gold. Chalcopyrite, pyrite, and pyrrhotite are all compositionally ideal, with only trace amounts of Bi, Zn and Ni (Table 2).

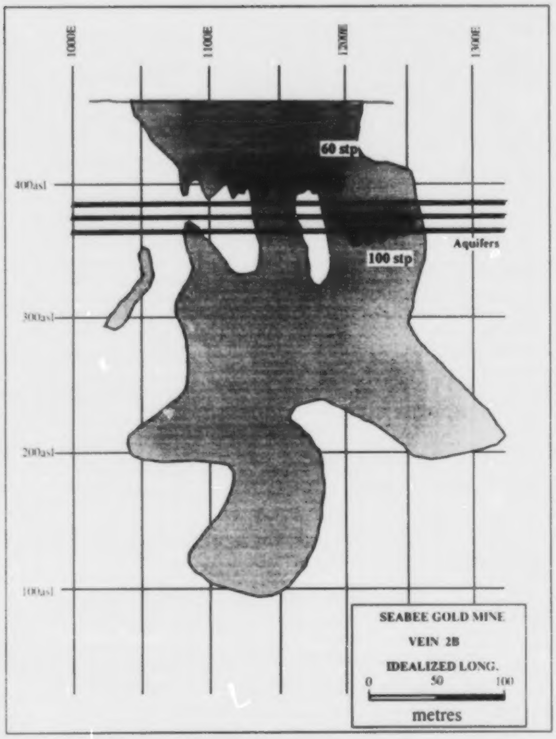


Figure 4 - Longitudinal section of orebody showing orientation of late flat fractures and locations of active stopes.

Free gold, ranging from 1 to 50 μm, occurs as disseminated grains in the veins, or as isolated blebs within other sulphide phases. Gold apparently co-precipitated with Bi-telluride or Fe-telluride (McLeod, 1982). The gold particles are of high tenor with less than 5 percent silver. Electron microprobe analyses yielded a composition of approximately $Au_{92}Ag_5$ for gold. Minor amounts of Se were detected in addition to trace concentrations of Bi in the 100-1000 ppm range.

Tellurides are in textural equilibrium with free gold and apparently these phases co-precipitated. Optically, the tellurides are spatially associated with, or in contact with, individual gold grains.

5. Stable Isotope Geochemistry

Stable isotope geochemistry was utilized to constrain the physical conditions of ore formation and the origin of the ore forming fluids.

Table 2 - Representative Microprobe Sulphide and Tourmaline Analyses (20 Kv and 15 Kv respectively).

Sulphide Mineral Analyses

Zone 2 S109-644.5' Auriferous Quartz Vein

Mineral	Po	Cpy	Py	Cpy	Py	Py	Po	Py	BiTe	Au	BiTe	Au
S	54.0	35.2	39.0	35.0	39.2	37.4	53.8	37.3	0.07	0.07	0.04	0.07
Fe	46.6	30.0	60.3	30.1	60.3	62.3	47.9	58.3	0.16	0.17	0.30	0.56
Ni	0.05	0.01	0.11	0.02	0.11	0.04	0.02	0.10	0.04	0.02	0.01	0.01
Cu	0.01	34.0	0.01	34.2	0.02	0.01	0.01	0.01	0.01	0.04	0.01	0.03
Zn	0.06	0.01	0.01	0.01	0.04	0.05	0.01	0.02	0.04	0.01	0.04	0.02
As	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.02	0.02	0.01	0.01
Se	0.01	0.01	0.13	0.01	0.01	0.01	0.01	0.09	0.01	1.50	0.01	1.92
Ag	0.01	0.01	0.01	0.05	0.01	0.01	0.01	0.01	0.01	3.06	0.01	4.01
Sb	0.01	0.01	0.01	0.01	0.02	0.01	0.01	0.01	0.18	0.02	0.03	0.01
Te	0.02	0.02	0.07	0.03	0.04	0.01	0.01	0.01	47.9	10.8	48.1	0.03
Au	0.01	0.01	0.01	0.01	0.01	0.88	0.01	0.01	0.55	74.6	0.62	93.3
Pb	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01
Bi	0.13	0.15	0.08	0.06	0.09	0.18	0.16	0.14	48.1	15.3	48.7	0.77
TOTAL	100.8	99.5	99.7	99.4	99.8	100.5	101.8	96.0	97.0	105.5	97.8	100.7

Cpy = Chalcopyrite, Py = Pyrite, Po = Pyrrhotite, Au = Native Gold, BiTe = Tellurobismuthite

Tourmaline Analyses

Zone 2 Sample - #8a Quartz Vein

Point #	PT#1	PT#2	PT#3	PT#4	PT#5	PT#6	PT#7	PT#8	PT#9	PT#10	PT#11
SiO_2	35.46	35.75	35.07	35.33	35.1	34.98	35.45	35.41	35.32	35.69	34.69
TiO_2	0.33	0.38	0.51	0.79	0.53	0.37	0.15	0.11	0.19	0.24	0.23
B_2O_3	10.27	10.35	10.16	10.23	10.17	10.13	10.27	10.26	10.23	10.34	10.05
Al_2O_3	30.00	29.93	29.79	29.70	28.23	29.13	28.55	30.36	29.05	29.14	28.29
FeO	8.83	8.92	9.44	8.94	9.37	8.89	9.04	9.14	9.29	9.35	9.19
MnO	0.00	0.03	0.03	0.02	0.03	0.00	0.05	0.00	0.01	0.00	0.00
MgO	7.17	7.44	6.45	6.75	7.90	7.06	7.30	6.72	7.48	7.27	7.16
CaO	0.45	1.02	0.39	0.77	1.08	0.48	0.33	0.18	0.29	0.28	0.30
Na_2O	2.24	1.95	2.19	2.04	2.20	2.23	2.40	2.12	2.42	2.37	2.28
K_2O	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
F	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
TOTAL	94.75	95.77	94.03	94.57	94.62	93.27	93.54	94.30	94.28	94.68	92.19

Reported data is in weight percent oxides for silicates and total elemental percent for sulfides.

Petrographic analysis of vein material was undertaken to identify suitable mineral pairs that co-precipitated in equilibrium during vein genesis (i.e. quartz-tourmaline-chlorite or pyrite-gold-telluride). Five surface quartz vein samples (V2a-V2b) and five drill core quartz samples which included a quartz-tourmaline-chlorite sample from DDH#S-126 (299.5 m), a tourmaline-quartz sample from S-133 (281.0 m) and quartz separates were analyzed for $\delta^{34}S$, $\delta^{18}O$, and δD as appropriate. Estimates of ambient vein temperatures, and of the $\delta^{18}O$ and δD for the water from which the minerals precipitated were calculated using accepted isotopic fractionation factors (see Wenner and Taylor, 1971; Kyser, 1987).

The detailed petrographic analysis established that auriferous veins contain quartz-tourmaline-chlorite-actinolite-carbonate in textural equilibrium. These phases apparently co-precipitated with gold. Quartz $\delta^{18}O$ values from both veins and drill core fall in a narrow range from 10.2 to 9.0 per mil (Table 3). Ambient temperatures of 360° to 510°C were determined for quartz-tourmaline and quartz-chlorite mineral pairs (Table 3) and, hence, gold mineralization probably occurred within this temperature range. Based on these temperature estimates, and mineral $\delta^{18}O$ values, a calculated $\delta^{18}O$ of 5.4 to 6.8 per mil was determined for the auriferous fluid.

In the Flin Flon domain there are considerable variations in $\delta^{18}O$ values (10.3 to 15.5) of unmineralized veins (Kyser, *et al.*, 1986; Ansdell and Kyser 1990; Fedorowich *et al.* in press), whereas mineralized veins at Tartan Lake show a more restricted range between 10.6 to 12.0 per mil (Fedorowich *et al.*, in press). Consequently, Ansdell and Kyser (1990) and Fedorowich *et al.* (in press) suggested that different vein generations could not be distinguished on the basis of oxygen isotopic abundances alone. However, Kyser *et al.* (1986) found a marked contrast in $\delta^{18}O$ between barren (12.5 to 13.0 per mil) and auriferous veins (10 to 11 per mil) for the Star Lake deposit in the La Ronge Domain. More

Table 3 - Oxygen, Hydrogen and Sulphur Isotopic Compositions of Mineral Separates from the Seabee Gold Deposit. Calculated temperatures and isotopic compositions of ore forming fluids are also included.

Sample Number	Type	Location	$\delta^{18}O$ silicates			$\delta^{34}S$		δD		Temp °C	$\delta^{18}O$ water	δD water
			qtz	tourm	chlor	py	po	chl	tourm			
Z10-1	Qtz vein	Zone 10	9.7									
Z2-1	Qtz vein	Zone 2	10.0						-66			
Z14-2	Qtz vein	Zone 14	9.8									
Z2-9	Qtz vein	Zone 2	9.3									
Z5-2	Qtz vein	Zone 5	10.2									
S-133 281.0m	Qtz vein	Zone 2	9.8				3.5					
S-143 180.5m	Qtz vein	Zone 2	9.8	8.2						510° q-t	6.8	
S-124 243.6m	Qtz vein	Zone 2	9.5				3.6					
S-126 299.5m	Qtz vein	Zone 2	9.7	7.0	4.0	2.8		-60		360° q-t	5.4	
										460° q-chl	5.4 q-chl	-35
S-129 141.7m	Qtz vein	Zone 2				1.2						
Z2-10	MD	Zone 2	8.3									
Z2-P	FP	Zone 2	9.5									
Z5-1	M S	Zone 5	8.7									

Minerals: qtz = quartz, tourm = tourmaline, chlor = chlorite, py = pyrite, po = pyrhotite.
Lithologies: MD = Metadiorite, MS = Metasediment, FP = Feldspar Porphyry
Fractionation Factors: quartz-chlorite (q-chl), Wenner and Taylor (1971), quartz-tourmaline (q-t) based on the muscovite fractionation factors after King (1990). Isotope δ values reported in per mil.

oxygen isotopic data for barren and auriferous veins must be obtained for veins at the Seabee gold deposit, and for the Glennie Domain in general, to resolve whether or not regional fluids differ from ore-forming fluids.

Hydrogen isotope measurements were performed on chlorite (δD = -60 per mil) and tourmaline (δD = -66 per mil) mineral separates. The δD values of the ore fluid then were calculated (δD = -30 per mil), using the appropriate mineral water fractionation factors, in conjunction with estimated equilibrium temperatures obtained from oxygen isotope mineral pairs. Calculated $\delta^{18}O$ and δD values of the auriferous fluids plot within the field of metamorphic water and indicate a system with low water rock ratios.

Sulphur isotope measurements on vein pyrite and pyrrhotite yielded $\delta^{34}S$ values of 1.2 to 3.6 per mil with a median of 2.8 per mil. These values are consistent with either a direct magmatic source, or leaching of magmatic sulphide minerals. Sulphur isotope values obtained from Seabee mineralization range from 1.3 to 2.8 per mil. They are similar to $\delta^{34}S$ values of 0 to 3 per mil for Star Lake (Kyser *et al.*, 1986) and values of 1.4 to 3.1 per mil for pyrite in auriferous veins from Tartan Lake (Fedorowich *et al.*, in press). The tight clustering of sulphur isotope values is a characteristic shared by Archean deposits and signifies a uniform sulphur source. Kerrich (1989) has argued that the restricted $\delta^{34}S$ values indicate that the auriferous fluids acquired sulphur from a uniform reservoir, and that the oxidation state was generally below the SO_2/H_2S buffer.

6. Conclusions

The principal conclusions that stem from research to date are as follows:

1) The Laonil Lake Intrusive Complex post-dates the early D_1 deformation event, and predates the development of the Ray Lake Synform (D_2). At peak or post-peak(?) lower amphibolite facies metamorphism and during emplacement of the Eyahpaise pluton, rheological contrasts within the LLIC facilitated the development of a shear zone system dominated by flattening or pure shear. The shear zones were the focus for emplacement of felsic intrusives and gold quartz-vein mineralization.

2) Mineralization and initial precipitation of gold appears to be syn-D_2 or to post-date the D_2 deformational event and could be associated with post-peak metamorphic conditions prior to the D_3 deformational event. Subsequently, during retrograde metamorphism, D_2 shear systems may have been reactivated during the D_3, resulting in remobilization or reprecipitation of earlier components, including gold, into later D_3 fracture systems.

3) Tourmaline compositions are in the schorl-dravite solid solution series, similar to those of other mesothermal gold deposits in the Proterozoic Trans-Hudson Orogen, and also to Archean counterparts.

4) Ore forming fluids were characterized by temperatures of 360° to 510°C and, on the basis of O and H isotopes, were likely derived from a metamorphic source undergoing devolatilization at the time of mineralization.

5) Slightly positive $\delta^{34}S$ values indicate that S was derived from an igneous source. Metamorphic devolatilization reactions could have remobilized sulphur, and consequently gold, from a source that was originally dominated by magmatic rocks. Conceivably, Au and sulphur may have been derived from the felsic porphyries associated with the

deposit, or alteratively, the ore-forming systems may have overprinted these porphyries.

6) The Seabee deposit has similar geologic setting, parageneses, temperature and ore-fluid chemistry as other mesothermal deposits in the La Ronge and Flin Flon Domains.

7. Acknowledgments

The senior author would like to thank Saskatchewan Energy and Mines for financial support during the past year, and, in particular, Tom Sibbald who provided help both in Saskatoon and in the field. In addition, the authors would like to express their appreciation to Gary Billingsley and Peter Gummer, of Claude Resources and to all the geologists at Seabee for extensive tours underground, technical information provided, and for counter arguments against geological interpretations presented by the author.

8. References

Ansdell, K.M. and Kyser, T.K. (1990): Epigenetic gold mineralization in the Flin Flon Domain: fluid characteristics; *in* Beck, L.S. and Harper, C.T. (eds.), Modern Exploration Techniques, Sask. Geol. Soc., Spec. Publ. No.10, p219-231.

Bickford, M.E., Collerson, K.D., Lewry, J.F., Van Schmus, W.R. and Chiarenzelli, J.R. (1990): Proterozoic collisional tectonism in the Trans-Hudson Orogen, Saskatchewan; Geol., v18, p14-18.

Bickford, M.E. and Van Schmus, W.R. (1985): Preliminary U-Pb age data for the Trans-Hudson Orogen in northern Saskatchewan: new and revised results; *in* Summary of Investigations 1985, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 85-4, p63-66.

Burnsall, J.F. (1989): Introduction: Review of the mechanical principles, deformation mechanisms and shear zone rocks; *in* Burnsall J.F. (ed.), Mineralization and Shear Zones, Geol. Assoc. Can., Short Course Notes, v6, p1-26.

Cambell, S.W. (1986): Petrographic Study of Samples from Seabee; Placer Development Limited, internal report.

Chiarenzelli, J.R. (1989): The Nistowiak and Guncoat Gneisses: implications for tectonics of the Glennie and La Ronge Domains, northern Saskatchewan, Canada; unpubl. Ph.D. Thesis, Univ. Kansas, 229p.

Clayton, R.N., (1963): The use of brominepentaflouride in the extraction of oxygen from oxides and silicates for isotopic analyses, Geochem. Cosmochim. Acta, v27, 43-52p.

Coombe, W. (1984): Gold in Saskatchewan; Sask. Energy Mines, Open File Rep. 84-1, 134p.

Delaney, G.D. (1986): Bedrock geological mapping, Laonil Lake area; *in* Summary of Investigations 1986, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 86-4, p32-41.

__________ (1991): Gold in the Glennie Domain, Lower Proterozoic Trans-Hudson Orogen, northern Saskatchewan; Geological Society of CIM, First Annual Field Conference, Saskatoon, Sept. 1991, Prog. Abstr., p11.

Delaney, G.D., Carr, S.D. and Parrish, R.R. (1988): Two U-Pb zircon ages from the eastern Glennie Lake Domain, Trans-Hudson Orogen, Saskatchewan; *in* Radiogenic Age and Isotope Studies, Geol. Surv. Can., Pap. 88-2, p51-58.

Delaney, G.D., Heaman, L.M., Kamo, S., Parrish, R.R., Slimmon, W.L. and Reilly, B.A. (1990): U-Pb sphene/zircon geochronological investigations: *in* Summary of Investigations 1990, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 90-4, p54-57.

Fedorowich, J., Stauffer, M., and Kerrich., R. (in press): Structural setting and fluid characteristics of the Proterozic Tartan lake gold deposit, Trans-Hudson Orogen, Northern Manitoba; Econ. Geol.

Helmstaedt, H. (1986): Report on the geology of the Seabee and Curie Rose Properties, Laonil Lake, Saskatchewan; Placer Development Limited, internal report.

__________ (1987): Report on the geology of the No.2 Zone on the Seabee Property, Laonil Lake, Saskatchewan; Placer Development Limited, internal report.

Kerrich, R. (1989): Geochemical evidence on the source of fluids and solutes for shear zone hosted mesothermal Au deposits; *in* Bursnall J.T. (ed.), Mineralization and Shear Zones, Geol. Assoc. Can., Short Course Notes, v6, p129-194.

King, R. (1990): Tourmaline from Mesothermal Gold Deposits of the Superior Province, Canada; Textural, Chemical and Isotopic relationships; unpubl. Ph.D. thesis, Univ. of Sask., 228p.

Kyser, T.K. (1987): Equilibrium fractionation factors for stable isotopes; *in* Kyser T.K. (ed.), Stable Isotope Geochemistry of Low Temperature Fluids, Mineral. Assoc. Can., Short Course v13, p1-84.

Kyser, T.K., Janser, B.W., Wilson, M.R. and Hattie, I. (1986): Stable isotope geochemistry related to gold mineralization and exploration in the Western Shield; in Clark L.A. (ed.), Gold in the Western Shield, CIM, Spec. Vol. 38, p470-498.

Lewry, J.F.(1977): The geology of the Glennie Lake area; Sask. Dep. Miner. Resour., Rep. 143, 59p.

__________ (1981): Lower Proterozoic arc-microcontinent collisional tectonics in the western Churchill Province; Nature, v294 no.5836, p69-72.

Lewry, J.F. and Collerson, K.D. (1991): The Trans-Hudson Orogen: Extent, subdivision, and problems; *in* Lewry, J.F. and Stauffer, M.R., (eds.), The Early Proterozoic Trans-Hudson Orogen of North America; Geol. Assoc. Can., Spec. Pap. 37, p1-14.

Lewry, J.F. and Sibbald, T.I.I. (1977): Variation in lithology and tectonometamorphic relationships in the Precambrian basement of northern Saskatchewan; Can. J. Earth Sci., v14, p1453-1467.

Lister, G.S. and Snokes, A.W. (1984): S-C Mylonites; J. Struc. Geol., v6, no.6, p617-638.

McLeod, J.A. (1983): Seabee petrographic study (Job V82:697R); Cominco Limited, internal report.

Poulsen, K.H. (1989): La Ronge structural studies; *in* Galley, A.G. (ed.), Investigations by the Geological Survey of Canada in Manitoba and Saskatchewan during 1984-89 Mineral Development Agreement; Geol. Surv. Can., Misc. Rep. 2133, p50-63.

Schultz, D.J. (1990): Reconnaissance geological study of the Seabee gold deposit, Glennie Domain; *in* Summary of Investigations 1990, Saskatchewan Geological Survey; Sask. Energy Mines, Misc. Rep. 90-4, p70-73.

Van Schmus, W.R. and Bickford, M.E.(1984): Preliminary age data for the Trans-Hudson Orogen in northern Saskatchewan; *in* Summary of Investigations 1984, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 85-4, p63-66.

Van Schums, W.R. Bickford, M.E., Lewry, J.F. and Macdonald, R. (1987): U-Pb zircon geochronology in the Trans-Hudson Orogen in northern Saskatchewan, Canada; Can. J. Earth Sci., v24, p407-424.

Wenner, D.B. and Taylor, H.P. (1971): Temperatures of serpentinization of ultramafic rocks based on $^{18}O/^{16}O$ fractionation between coexisting serpentine and magnetite; Contrib. Mineral. Petrol., v32, p165-185.

Mineralization and Associated Alteration in the Hanson Lake Area[1]

R.O. Maxeiner[2] and B.R. Watters[2]

Maxeiner, R.O. and Watters, B.R. (1991): Mineralization and associated alteration in the Hanson Lake area; *in* Summary of Investigations 1991, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 91-4.

The present study is aimed at improving the understanding of the relationship between supracrustal rocks and associated alteration and base metal mineralization in the western Hanson Lake area (Figure 1). It is part of an M.Sc. thesis project being conducted at the University of Regina in conjunction with the Saskatchewan Geological Survey. Altered rock sequences and associated base metal mineralization in two locations (Figure 1) were mapped at 1:10 000 scale and sampled in detail. A suite of 117 samples, collected by T.I.I. Sibbald during the summers of 1989 and 1990, provided a basis for petrographic and geochemical examination of the major rock types. Hence, this report is based on a combination of petrographic and geochemical work done on samples collected during the last two years and new field work accomplished during the past summer.

1. Previous Work

The area was first mapped at a scale of 1:126,720 by Wright and Stockwell (1934) for the Geological Survey of Canada. Byers (1957) carried out a more detailed investigation at a scale of 1:31,680 for the Saskatchewan Department of Mineral Resources. Both the geology and geochemistry of the Hanson Lake area were examined by Coleman *et al.* (1970), who produced a map at a scale of 1:15,840. Lithogeochemical studies have also been undertaken by Parslow and Gaskarth (1986). The most recent work was carried out by Sibbald (1989), who mapped the area of western Hanson Lake at 1:20 000 scale.

2. Regional Geology

The Hanson Lake area lies within the Hanson Lake Block in the southeastern part of the Reindeer Zone of the Early Proterozoic Trans-Hudson Orogen. The Hanson Lake Block comprises elements of both Archean and Early Proterozoic age, the former represented by the charnockitic Sahli Granite and the MacMillan Point Granite. A U-Pb age of 2500 Ma has been reported for the Sahli Granite (Lewry *et al.*, 1987). These Archean cores are thought to have been overridden by nappe sheets during the Early Proterozoic collision with the Rae-Hearne palaeoplatform (Lewry *et al.*, 1990; Hoffman, 1990). The supracrustal rocks of the block are of Early Proterozoic age and have features in common with the Flin Flon belt which has been interpreted to represent an early stage island arc (Syme, 1990). A quartz-feldspar porphyry (previously interpreted as rhyolite), collected at the Western Nuclear Mine, yielded an age of 1888 ±12 Ma (Lewry *et al.*, 1987).

The area of investigation includes both mafic and felsic Proterozoic volcanic and sedimentary rock units that lie in the southeast corner of the Hanson Lake Block, just north of the Phanerozoic dolomite cover. The supracrustal units structurally overlie the Jackpine and Tulabi Lakes anticlines situated to the northwest of the area studied. The anticline cores dominantly comprise granodiorites and migmatites. To the east, the Sturgeonweir thrust separates the supracrustal rocks from similar rocks of the Amisk Group (Ashton *et al.*, 1987). Low grade Amisk Group rocks of the Flin Flon Belt structurally overlie higher grade supracrustal rocks of the Hanson Lake Block.

Within the area of study, the metamorphic grade increases from lower amphibolite in the south to middle amphibolite in the north. Interpretation of the structural geology is inhibited because primary compositional layering is only clearly seen in some of the graywacke assemblages. Where observed it strikes parallel to the north-south-trending foliation. No repetition of major rock types has been noted in the area mapped.

Two significant base metal deposits are known in the Hanson Lake area. The Western Nuclear Mine is an old Pb-Zn producer, that was in operation from 1967-69 and produced 147 300 tonnes of ore, with grades of 5.83 percent Pb and 9.99 percent Zn. Silver was recovered as a by-product. The McIlvenna Bay deposit is currently under investigation and has estimated reserves of at least 10.7 million tonnes of ore, with grades of 5.4 percent Zn and 1 percent Cu with accessory Au and Ag.

3. Description of Major Rock Types

a) Dacite

Dacite is the main volcanic rock type in the Hanson Lake area. It is exposed in the vicinity of the former Western Nuclear Mine and, because the sequences are believed to young eastwards, represents the lowest stratigraphic unit mapped. The dacitic units are very fine-grained, dark grey rocks. Primary features are not

---

(1) Saskatchewan Project A.146, from which this report derives, was funded in 1991 under the Canada-Saskatchewan Partnership Agreement on Mineral Development 1990-95.

(2) Department of Geology, University of Regina, Regina, Saskatchewan

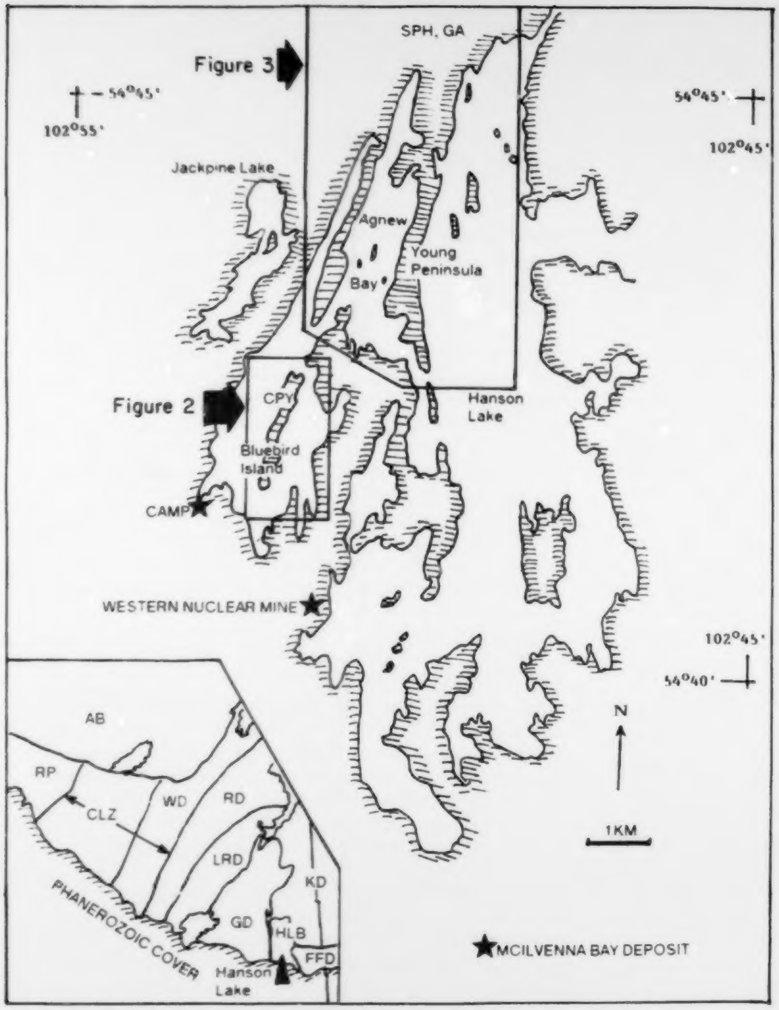


Figure 1 - Location of the western Hanson Lake area. Inset: arrow marks the location of the Hanson Lake area within the Trans-Hudson Orogen of northern Saskatchewan; AB = Athabasca Basin; HLB = Hanson Lake Block; FFD = Flin Flon Domain; KD = Kisseynew Domain; GD = Glennie Domain; LRD = La Ronge Domain; RD = Rottenstone Domain; WD = Wollaston Domain; CLZ = Cree Lake Zone; RP = Rae Province. Main figure: CPY = chalcopyrite occurrence; SPH, GA = sphalerite and galena occurrences.

generally preserved, but it is believed that the dacitic units represent a sequence of monolithologic flow breccias. A representative chemical analysis of dacite from Bluebird Island is given in Table 1.

On Bluebird Island (Figure 2) the dacite is garnet-bearing (5 percent) with the garnets surrounded by biotite rims. Elongate biotite lenses are scattered throughout the rock and are in part streaked out into the foliation. The rocks are moderately foliated and the mafic content ranges from 5 to 10 percent. In thin-section the dacite has a micro-porphyritic texture with plagioclase and subordinate microcline phenocrysts, which are sometimes difficult to recognize because they are extensively saussuritized. The phenocrysts occur in a groundmass of plagioclase, quartz and biotite. Some of the plagioclase crystals are zoned. The main constituents of the dacite, in decreasing order of abundance, are; plagioclase, quartz, biotite, microcline, muscovite and garnet; accessory minerals are epidote, carbonate, apatite, tourmaline and opaques.

b) Garnet-Sericite Schist

This newly recognized unit occurs in the northern part of Bluebird Island in contact with highly altered rock units and in close proximity to sulphide mineralization. The unit is very well foliated and contains abundant, rounded, pink garnet crystals, that are contained in a very fine-grained, grey, sericitic, and slightly silicified matrix. The garnet content varies from 10 to 15 percent and garnet crystals are typically 1 to 5 mm in size, commonly situated within elongate biotite lenses. The garnet-sericite schist is distinctly different in appearance to the dacite, but is marginally intercalated with garnetiferous dacite. Field evidence suggests that the garnet-sericite schist might represent altered dacite that has been sericitized and garnetized. Microscopically, these rocks contain large poikiloblastic garnet crystals set in a fine-grained quartz-sericite groundmass. Feldspar is conspicuously rare or absent.

c) Garnet-Amphibole Assemblage

The garnet-amphibole assemblage has also been mapped in association with chalcopyrite mineralization in the northern part of Bluebird Island (Figure 2). The rocks are characterized by amphibole veins and lenses that are dominantly oriented parallel to the foliation, but are also found crosscutting it and crosscutting each other. The amphibole content of the assemblage averages around 15 percent and the garnet content ranges from 5 to 10 percent. In outcrop the rocks appear massive and the immediate host to the amphibole veins is a silicified and slightly carbonatized, fine-grained garnetiferous dacite. The amphibole veins are typically 0.5 to 2 cm wide and irregularly spaced. Garnet-amphibole rocks occur in direct contact with malachite and chalcopyrite mineralization in trenches on the northern part of the island and are, therefore, tentatively interpreted to be part of a footwall alteration zone, representing a metamorphosed and deformed chlorite alteration assemblage.

Table 1 - Representative Geochemical Analyses of Major Rock Types from the Hanson Lake Area. Major elements in percent, trace elements in ppm.

Sample #	9050-35	8950-09	9050-68	9050-66	9050-04	9050-67	9050-53	9050-41	8950-27	9050-32	9050-21
SiO_2	53.25	62.28	69.25	56.59	76.77	47.13	78.39	65.58	59.04	71.79	66.08
TiO_2	0.56	0.47	0.35	0.56	0.14	0.41	0.28	0.26	0.71	0.28	0.39
Al_2O_3	11.33	15.12	14.07	10.21	14.11	1.83	11.74	13.66	14.72	13.62	15.92
FeO_{tot}	8.59	3.92	6.38	12.37	0.17	33.2	3.77	2.56	5.91	3.32	3.18
MnO	0.19	0.08	1.77	2.13	0.21	6.2	0.12	0.06	0.09	0.14	0.05
MgO	11.98	2.54	1.38	7.22	0.12	7.72	2.37	6.89	5.63	0.38	1.66
CaO	5.19	4.05	0.27	5.47	0.08	2.01	0.15	7.1	4.85	2.19	2.94
Na_2O	1.27	3.30	0.12	0.55	0.77	0.02	0.12	1.26	3.91	3.57	4.87
K_2O	3.56	3.06	4.41	2.43	7.43	0.02	2.13	1.28	1.98	2.62	2.24
P_2O_5	0.23	0.13	0.07	0.27	0.02	0.06	0.07	0.05	0.36	0.05	0.11
H_2O^+	0.07	0.07	0.06	0.00	0.07	0.00	0.00	0.00	0.00	0.00	0.07
L.O.I.	1.6	2.63	0.99	1.10	0.39	0.59	1.39	1.73	1.45	1.06	0.93
Total	97.82	97.65	99.06	98.96	100.28	99.26	100.53	100.43	98.65	99.02	98.44
Zn	370	50	92	769	3	514	130	84	74	62	37
Cu	59	2	152	nd	54	nd	1	18	nd	nd	1
Ni	317	7	2	165	nd	66	15	nd	34	nd	2
Cr	751	nd	nd	736	nd	503	nd	nd	61	nd	nd
V	110	729	97	4	106	7	5	122	5	nd	54
Ba	789	670	235	356	990	18	108	404	705	822	585
Nb	10	7	12	10	15	4	11	25	9	23	7
Zr	98	98	139	95	126	35	115	261	157	232	92
Y	19	7	32	20	14	27	17	66	19	57	6
Sr	149	504	15	157	43	2	6	261	1002	178	591
Rb	68	42	69	54	90	4	38	26	23	45	33

9050-35 mafic volcanic
8950-09 dacite
9050-68 garnet-sericite schist
9050-32 quartz-feldspar porphyry (Bluebird Island)
9050-21 quartz-feldspar porphyry (Young Peninsula)
nd = not determined

9050-66 garnet-amphibole assemblage
9050-04 cherty rhyolite
9050-67 banded iron formation
9050-53 garnet-sillimanite schist
9050-41 calc-silicate
8950-27 micro-diorite

On average, chalcopyrite makes up less than 1 percent of the overall assemblage, but can be up to 5 percent in small, heavily mineralized pods. Where visible, chalcopyrite occurs as very small veinlets crosscutting the weakly developed schistosity, a feature consistent with interpretation of the unit as a footwall alteration assemblage. A sample from one of the trenches yielded 1.57 percent Cu and 0.23 percent Zn. In thin-sections the amphibole is hornblende and actinolite occurring as veinlets closely intergrown with garnet and opaque minerals. The fine-grained matrix is mainly composed of quartz and biotite. In this unit also, feldspar is conspicuously absent or completely altered to sericite.

d) Mixed Banded Iron Formation (Silicate-Oxide-Sulphide Facies)

A magnetite-bearing, mixed silicate-oxide-sulphide facies banded iron formation (BIF) can be traced over the northern part of Bluebird Island. It is fairly well exposed in two small trenches and is characterized by a subtle layering. Magnetite-bearing parts are intercalated with amphibole-garnet-bearing layers. The unit stratigraphically overlies the sulphide mineralization and the associated amphibole stringer rocks; however, it appears also partly overlain by the latter.

Banded iron formation also constitutes part of the upper sedimentary assemblage on Young Peninsula (Figure 3). Although apparently situated considerably higher stratigraphically, this iron formation is very similar to the iron formation on Bluebird Island.

Thin section examination indicates that the iron formation contains amphibole (cummingtonite-grunerite), magnetite, extremely fine-grained quartz and garnet. The garnet is anhedral, poikiloblastic and patchy. Concentrations of the different minerals produces layering on a centimetre scale. Chemically, the rocks of the iron formation are low in Al_2O_3 and rich in Fe and Mn (Table 1).

e) Mafic Volcanic Unit

Very indistinct, dark grey to black, fine-grained biotite-amphibole rocks, outcropping only along the northeast shore of Bluebird Island and lacking in primary textures, have been tentatively interpreted as basalt. In thin section the rocks appear moderately foliated with abundant brown biotite defining the schistosity. Disrupted poikiloblastic hornblende phenocrysts are still visible and distinct from post-kinematic finer-grained hornblende.

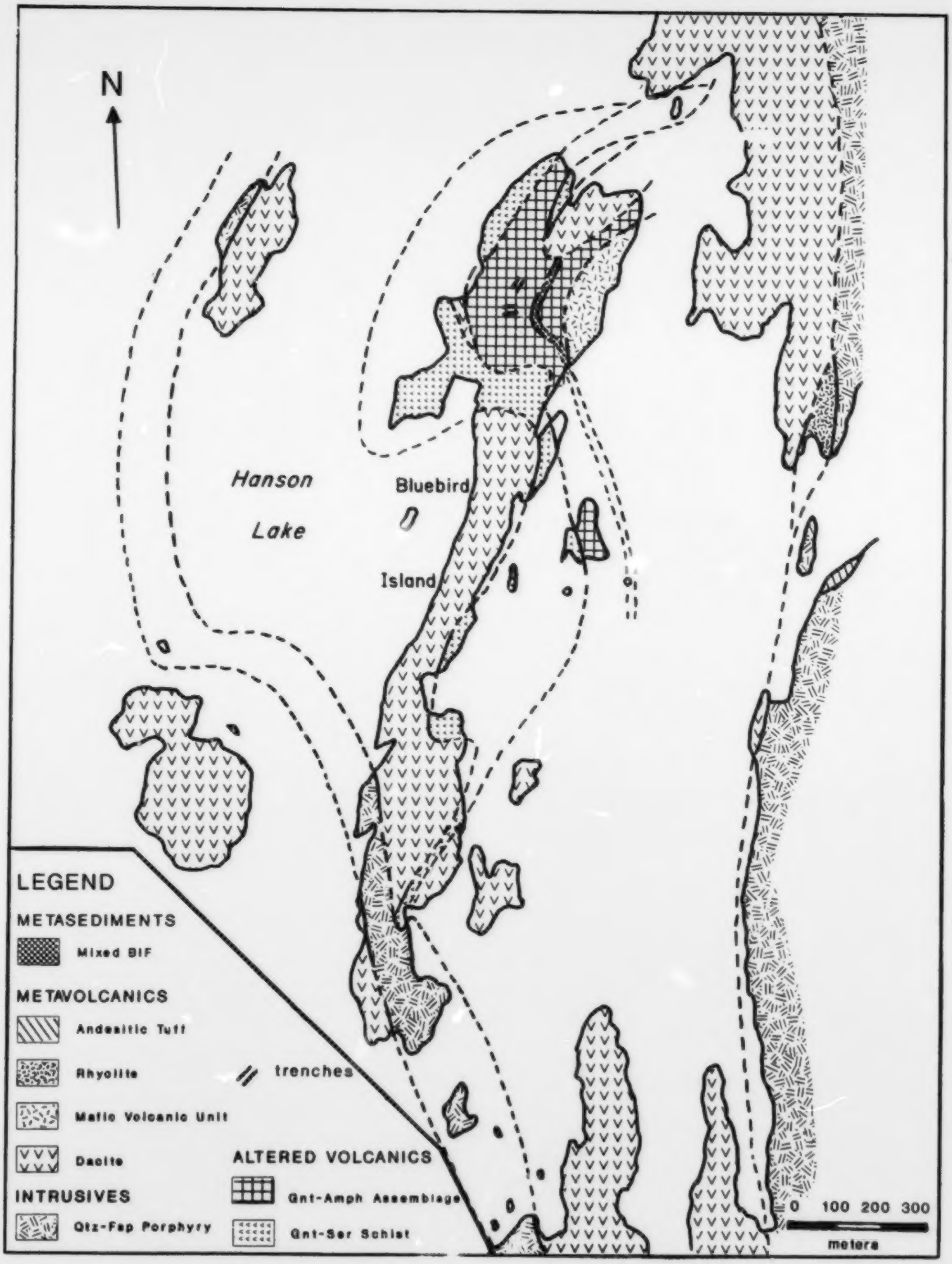

N
Hanson Lake
Bluebird Island
trenches
LEGEND
METASEDIMENTS
Mixed BIF
METAVOLCANICS
Andesitic Tuff
Rhyolite
Mafic Volcanic Unit
Dacite
INTRUSIVES
Qtz-Fsp Porphyry
ALTERED VOLCANICS
Gnt-Amph Assemblage
Gnt-Ser Schist
0 100 200 300
meters

Figure 2 - Detailed geological sketch map of Bluebird Island. BIF = banded iron formation; Qtz-Fsp = quartz-feldspar; Gnt-Amph = garnet-amphibole; Gnt-Ser = garnet-sericite.

f) Lower Sedimentary Assemblage

The lower sedimentary assemblage strikes along the northeast shore of Agnew Bay and can be traced over a length of almost 2 km. It is a composite assemblage that hosts sphalerite and galena showings, and there are trenches in two locations. The sequence comprises garnet-mica schist, sillimanite schist, garnet-amphibole-diopside gneiss, massive ultramafic amphibolite and cherty fuchsite schist. Due to the extreme heterogeneity of the assemblage, and a lack of traceable sub-units, it has here been incorporated into one mappable unit. Layers within the sequence appear to pinch and swell along strike: large scale boudinage features are well documented in the northern trench, where a sillimanite-rich layer greater than 25 m thick pinches down to 5 m. In this trench the concordant subtle layering within the cherty garnet-amphibole-diopside gneiss indicates that this rock is lean sedimentary iron formation as opposed to discordant alteration. Intercalated with mica and sillimanite schists, it suggests deposition at low sedimentation rates in a low energy, subaqueous environment. Facies variations and tectonic boudinaging have led to the present disrupted appearance of the sequence. Mineralization exposed in the trenches is dominantly in the cherty, fuchsite-bearing parts of the assemblage and consists of sphalerite and galena, both as small veinlets and disseminations. In the southern trench on Young Peninsula, sphalerite occurs as 1 to 5 cm wide veins within highly silicified garnet-amphibole gneiss. Galena is visible in smaller blebs. Some of the amphibole in this trench is anthophyllite, a mineral that has been described in association with metamorphosed massive sulphide deposits (Friesen *et al.*, 1982). Subtle layering in these units and the close association with muddy sequences suggests the mineralization represents a distal exhalative horizon.

g) Rhyolite

A large body of cherty rhyolite is exposed on the islands in the southern part of Agnew Bay and along some of the surrounding lakeshores (Figure 3). The rocks are massive, cherty and extremely fine-grained, white to pink, and show a very delicate lamination on a millimetre-scale, possibly representing flow-banding. In thin section the rhyolites have an equigranular mosaic texture of fine-grained quartz, K-feldspar and minor plagioclase. Muscovite occurs as small aligned flakes defining a weak foliation. The feldspar is lightly sericitized.

h) Clastic Rhyolite Apron

Quartz-eye schists occur in close association with the cherty rhyolites. They are generally highly foliated and have a high proportion of quartz-eyes, comprising about 20 to 30 percent. The high proportion of quartz-eyes and the close association with intercalated sillimanite schists suggest a sedimentary origin for these rocks, probably having been derived from adjacent rhyolitic domes. They are believed to be proximal deposits, as they have basically the same mineralogy as the cherty rhyolite and feldspar is still preserved. Within these clastic aprons adjacent to the rhyolitic domes, there are also a number of coarse rhyolitic breccias. Large, pink fragments, 5 to 10 cm in size and of cherty rhyolitic composition, "float" in an amphibole rich matrix.

i) Fine Amphibolite

Fine-grained amphibolite has been mapped in the northeastern part of Young Peninsula. It is fairly inhomogeneous on a large scale, with hornblende content, ranging from 10 to 30 percent. The rocks are moderately foliated, with hornblende defining a stretching lineation. A typical thin section contains short, prismatic, subhedral hornblende crystals and moderately saussuritized plagioclase, with only minor amounts of quartz, muscovite, epidote, sericite and opaque minerals. Although in individual outcrops these rocks appear fairly homogeneous with an apparent lack of primary textural features, it is believed that they are a wacke-type mafic sedimentary deposit because they are, in places, intercalated, on a metre-scale, with layered graywackes and sillimanite schists.

j) Upper Sedimentary Assemblage

This diverse assemblage stratigraphically overlies the amphibolite units but may also be partly intercalated with them. It includes calc-silicates, psammitic graywacke, sillimanite schist and impure marble (30 to 40 percent carbonate content), of which only the first two will be described in detail. In the northern part of the area the assemblage is intruded by large pegmatite bodies.

Calc-silicate rocks form a massive unit, characterized by high free carbonate content and massive yet pod-like appearance. In addition to the 5 to 10 percent carbonate, plagioclase, actinolite, diopside and clinozoisite are essential constituents and sphene, zircon, apatite, sericite and epidote are accessory minerals. Outcrops form a distinct ridge in the northeastern part of Young Peninsula. The rocks are fine to medium grained, inhomogeneous, with moderate to strong foliation, and rodding features. Local psammitic layers grade laterally into typical psammitic graywackes, in which layering and in places graded bedding are preserved. Hence, the whole unit is interpreted as sedimentary with variations in composition along strike representing changes in facies. Coleman *et al.* (1970) interpreted the unit as a mixture of volcanic material and calcareous sediments.

Psammitic graywackes occur as lateral extensions of the calc-silicate rocks and overlie the latter to the east. The graywackes are generally very well layered and folded in places. Graded bedding occurs in layered graywacke units along the east shore of Young Peninsula and indicates younging towards the east. Layering is preserved on a centimetre scale with quartzofeldspathic layers predominating over amphibole-rich layers and garnet-rich layers. A staurolite-sillimanite-garnet-biotite assemblage was observed in rocks on islands east of Young Peninsula. Very coarse, clast-supported, pebbly layers feature quartz-eyes and lithic "eyes" several centimetres in size and elongated at a ratio of 1:5.

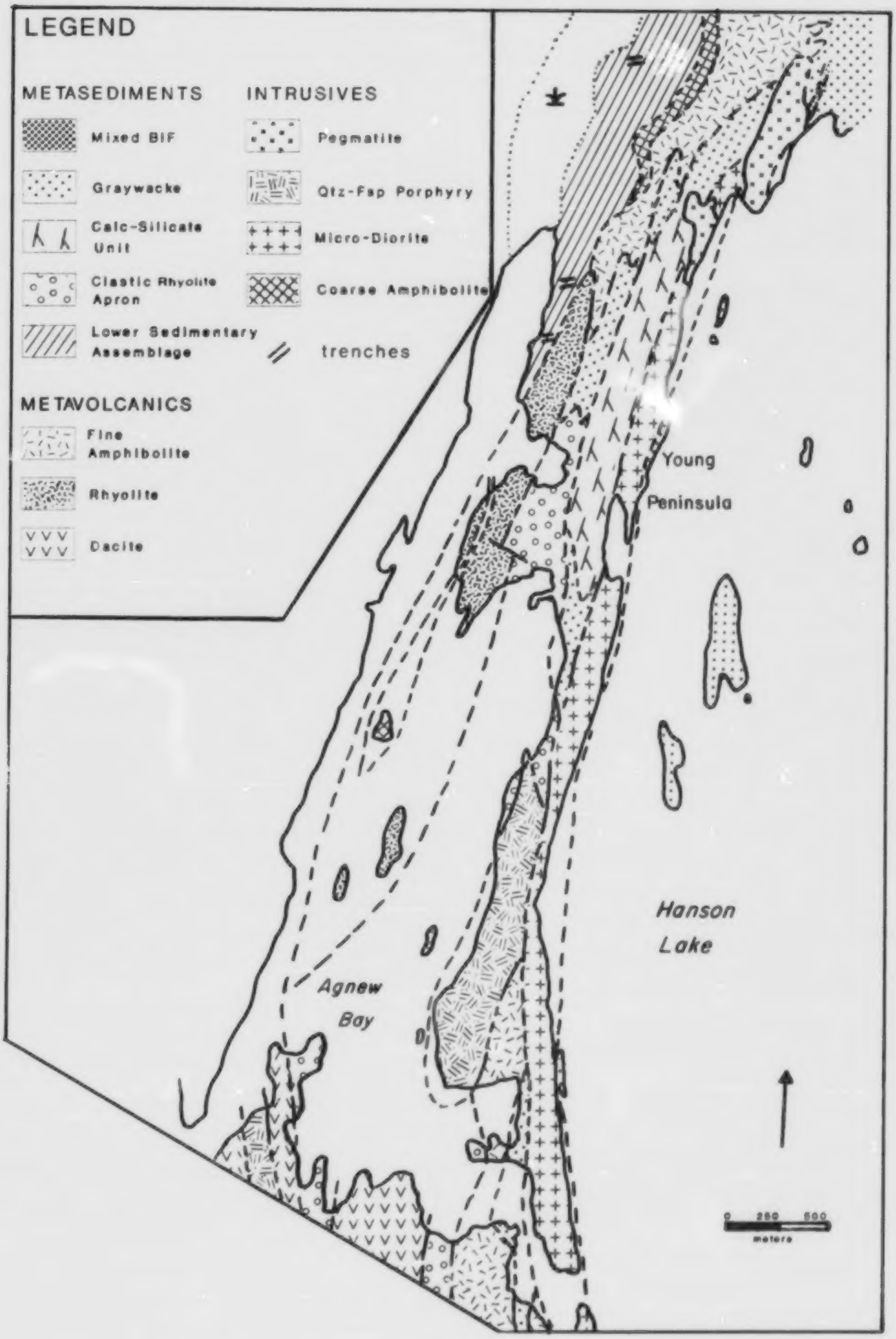

LEGEND
METASEDIMENTS
Mixed BIF
Graywacke
Calc-Silicate Unit
Clastic Rhyolite Apron
Lower Sedimentary Assemblage
METAVOLCANICS
Fine Amphibolite
Rhyolite
Dacite
INTRUSIVES
Pegmatite
Qtz-Fsp Porphyry
Micro-Diorite
Coarse Amphibolite
trenches
Young Peninsula
Hanson Lake
Agnew Bay
0 250 500
metres

Figure 3 - Detailed geological sketch map of Young Peninsula. BIF = banded iron formation; Qtz-Fsp = quartz-feldspar. The greywacke and calc-silicate units constitute the Upper Sedimentary Assemblage described in the text.

k) Quartz-feldspar Porphyry

Two bodies of quartz-feldspar porphyry have been mapped in detail. These rocks are massive and unfoliated. Eyes of quartz and pink feldspar crystals, 2 to 3 mm in size, give the rocks a porphyritic texture. Quartz eyes typically make up about 5 percent of the rock and are set in a fine-grained matrix. The mafic mineral content is also typically around 5 percent. The porphyry body on Bluebird Island contains up to 2 percent rounded garnet crystals which are up to 1 cm in size. The quartz-feldspar porphyry bodies are clearly intrusive in character and are possibly late subvolcanic phases.

l) Micro-diorite

The upper sedimentary assemblage has been intruded by a homogeneous, medium-grained, porphyritic, sill-like dioritic body interpreted by Coleman *et al.* (1970) as volcanic. In one location, large (20 to 50 cm) xenoliths of rhyolite provide evidence of its intrusive character. The rock-type contains abundant phenocrysts of plagioclase, 3 to 5 mm in size. Hornblende phenocrysts are relatively much scarcer. The body is extensive and can be traced from the north end of Young Peninsula to Sareco Island and farther south.

m) Coarse Amphibolite

A coarse-grained amphibolite can be traced as a distinct unit over the northeastern part of Young Peninsula and also occurs on an island in Agnew Bay. In one outcrop xenoliths of the rhyolite indicate the rock is intrusive. The amphibolite is massive and unfoliated and contains up to 50 percent hornblende phenocrysts up to 0.5 cm in size. Other major constituents are plagioclase and biotite.

4. Geochemistry

On a log Ti-log Zr plot (Figure 4), which is discriminatory for both fresh and altered rocks (Pearce *et al.*, 1981), the rhyolite, dacite, quartz-feldspar porphyry and micro-diorite units plot within the field of evolved island arc-type volcanics. Notably, each rock unit defines a distinct compositional grouping in terms of these two elements and, furthermore, the two quartz-feldspar porphyry intrusions are compositionally distinct from one another.

A SiO_2-log Zr/TiO_2 diagram (Figure 5) was used to classify the major rock groups and also shows the difference between two porphyry bodies, the one from Bluebird Island being almost rhyolitic in composition, the other being dacitic. This plot also confirms that rock types classified on the basis of field character as dacites are, in fact, dacitic in composition. Similarly, the micro-diorites are shown to be intermediate in composition.

A Jensen cation plot (Figure 6) indicates that the rhyolitic, dacitic and micro-dioritic units are calc-alkaline in nature. Once again, the two quartz-feldspar porphyry bodies display different compositional characteristics: the Young Peninsula body is calc-alkaline while the Bluebird Island body is relatively poor in Mg and, consequently, straddles the calc-alkaline-tholeiitic boundary.

5. Conclusions

Dacitic flow breccias, representing the major rock type on Bluebird Island (Figure 2), have been altered in various ways with two alteration zones recognized: an outer garnet-sericite zone and an inner garnet-amphibole zone, the latter hosting chalcopyrite mineralization and featuring faint stringer development. A preliminary interpretation envisages a scenario where the dacites have been penetrated by circulating hydrothermal fluids, causing alteration patterns similar to those described below volcanogenic massive sulphide deposits (Lydon, 1984 and 1988). Chemical and mineralogical evidence from samples taken on Bluebird Island supports this theory, as they are generally enriched in FeO, MgO and MnO and depleted in CaO and Na_2O relative to the dacitic rocks (Table 1). Feldspar is either highly altered or absent, which is to be expected in an environment of circulating hydrothermal fluids. Analysis of further samples, collected during the

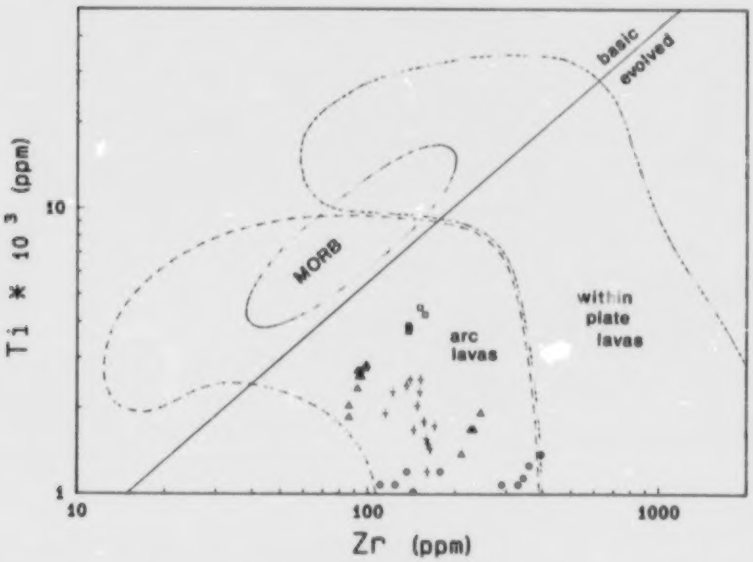


Figure 4 - Plot ot Ti vs. Zr for selected samples of the western Hanson Lake area. Field boundaries after Pearce and Norry (1979). Open circles, rhyolite; crosses, dacite; triangles, quartz-feldspar porphyry; squares, micro-diorite.

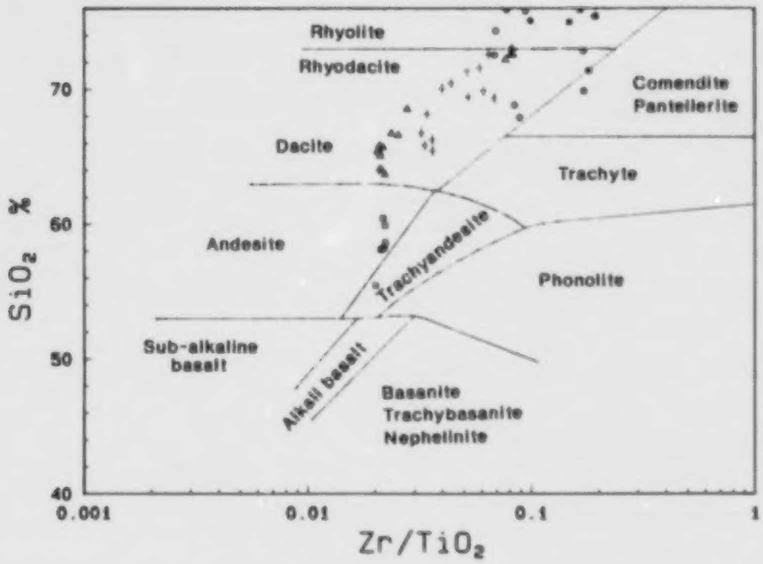


Figure 5 - Plot of SiO_2 vs. Zr/TiO_2 for selected samples of the western Hanson Lake area. Field boundaries after Winchester and Floyd (1977). Symbols as for Figure 4.

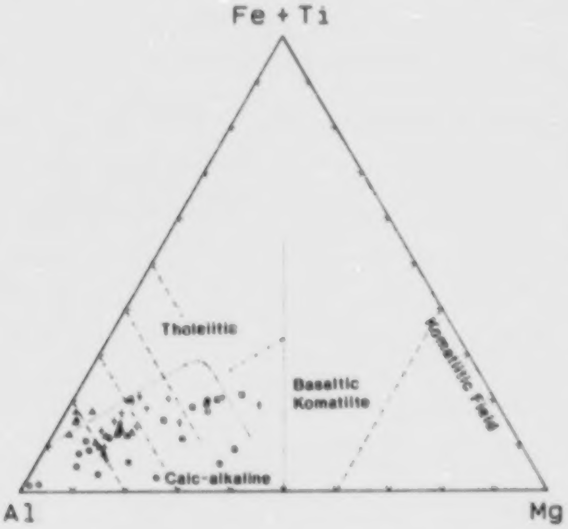


Figure 6 - Jensen cation plot for volcanic and subvolcanic rocks of the western Hanson Lake area. Field boundaries after Jensen (1976). Symbols as for Figure 4.

past summer, is being carried out to test this hypothesis.

Mineralization on Young Peninsula differs from that on Bluebird Island in that it is hosted by metasedimentary rocks and features sphalerite and galena development only; one feature in common with the two areas of mineralization is the association with garnet-amphibole-diopside assemblages. On Young Peninsula these rock types are believed to represent synsedimentary chemical precipitates, possibly formed in close proximity to hydrothermal feeder pipes. Sphalerite and galena are to be expected when hydrothermal fluids leach felsic volcanics and sediments (Lydon, 1988), whereas Cu-minerals will be generated in mafic to intermediate settings. Hence, the Pb-Zn showings of Young Peninsula also fit a model of small convective hydrothermal systems and could represent an exhalative horizon formed some distance from a central vent.

The preliminary geochemical data suggest generation of the volcanic and subvolcanic units of the western Hanson Lake area in a calc-alkaline island arc environment. The stratigraphically lowest quartz-feldspar porphyry body shows a transitional calc-alkaline to tholeiitic character. Earlier geochemical studies in the area (Parslow and Gaskarth, 1986), pointed out that the volcanic sequence is eastward facing with the more evolved rocks predominant in the east; the majority of their data were tholeiitic in character. This study basically supports the first observation, as rhyolitic rocks occur to the east of dacitic units, but differs in that the majority of the data point to a calc-alkaline island arc setting. Analyses of an additional 106 samples, collected during the past summer, are expected to shed more light onto the nature of the altered supracrustals from western Hanson Lake.

6. Acknowledgments

T.I.I. Sibbald was a great help during the first week of the summer. Discussions with T.I.I. Sibbald and D. Thomas during subsequent brief visits to the field were a great benefit. Many thanks to Anette, who provided great assistance in the field. T.I.I. Sibbald critically reviewed and upgraded an earlier version of this manuscript.

7. References

Ashton, K.E., Wilcox, K.J., Wheatley, K.J., Paul, D. and de Tombe, J. (1987): The boundary zone between the Flin Flon Domain, Kisseynew Gneisses and Hanson Lake Block, northern Saskatchewan; *in* Summary of Investigations 1987, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 87-4, p131-134.

Byers, A.R. (1957): The geology and mineral deposits of the Hanson Lake area; Sask. Dep. Miner. Resour., Rep. 30, 47p.

Coleman, L.C., Gaskarth, J.W. and Smith, J.R. (1970): Geology and geochemistry of the Hanson Lake area, Saskatchewan; Sask. Res. Counc., Rep. 10, 156p.

Friesen, R.G., Pierce, G.A. and Weeks, R.M. (1982): Geology of the Geco base metal deposit; *in* Hutchinson, R.W., Spence, C.D. and Franklin, J.M. (eds.), Precambrian sulphide deposits, Geol. Ass. Can., Spec. Pap. 25, p343-365.

Hoffman, P. (1990): Subdivision of the Churchill Province and extent of the Trans-Hudson Orogen; *in* Lewry, J.F. and Stauffer, M.R.(eds.), The Early Proterozoic Trans-Hudson Orogen of North America, Geol. Assoc. Can., Spec. Pap. 37, p15-39.

Jensen, L.S. (1976): A new cation plot for classifying subalkalic volcanic rocks; Ont. Geol. Surv., Misc. Pap. 66.

Lewry, J.F., Macdonald, R., Livesey, C., Meyer, M., Van Schmus, W.R. and Bickford, M.E. (1987): U-Pb geochronology of accreted terranes in the Trans-Hudson Orogen in northern Saskatchewan, Canada; *in* Pharach, T.C., Beckinsale, R.D. and Rickard, D. (eds.), Geochemistry and Mineralization of Proterozoic Volcanic Suites, Geol. Soc. Lon., Spec. Publ. 33, p147-166.

Lewry, J.F., Thomas, D.J., Macdonald, R. and Chiarenzelli, J. (1990): Structural relations in accreted terranes of the Trans-Hudson Orogen, Saskatchewan: telescoping in a collisional regime?; *in* Lewry, J.F. and Stauffer, M.R.(eds.), The Early Proterozoic Trans-Hudson Orogen of North America, Geol. Assoc. Can., Spec. Pap. 37, p75-94.

Lydon, J.W. (1984): Volcanogenic massive sulphide deposits Part I: a descriptive model; Geosci. Can., v11, p195-202.

__________ (1988): Volcanogenic massive sulphide deposits Part II: genetic models; Geosci. Can., v15, p43-65.

Parslow, G.R. and Gaskarth, J.W. (1986): Geochemistry of the Hanson Lake area; Sask. Energy Mines, Open File Rep. 86-1, 107p.

Pearce, J.A., Alabaster, T., Shelton, A.W. and Searle, M.P. (1981): The Oman ophiolite as a Cretaceous arc-basin complex: evidence and implications; Phil. Trans. Roy. Soc., A 300, p299-317.

Pearce, J.A. and Norry, M.A. (1979): Petrogenetic implications of Ti, Zr, Y and Nb variations in volcanic rocks; Contrib. Mineral. Petrol., v69, p33-47.

Sibbald, T.I.I. (1989): Base metal deposits and geology of the Early Proterozoic Hanson Lake Metavolcanics; *in* Summary of Investigations 1989, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 89-4, p66-70.

Syme, E.C. (1990): Stratigraphy and geochemistry of the Lynn Lake and Flin Flon metavolcanic belts; *in* Lewry, J.F. and Stauffer, M.R.(eds.), The Early Proterozoic Trans-Hudson Orogen of North America, Geol. Assoc. Can., Spec. Pap. 37, p143-161.

Winchester, J.A. and Floyd, P.A. (1977): Geochemical discrimination of different magma series and their differentiation products using immobile elements; Chem. Geol., v20, p325-343.

Wright, J.F. and Stockwell, C.H. (1934): West half of the Amisk Lake area, Saskatchewan; Geol. Surv. Can., Sum. Rep. 1933, Part C, p12-22.

Rare Element Pegmatites in the Hanson Lake Pegmatite Field[1]

D. G. MacDougall

MacDougall, D.G. (1991): Rare element pegmatites in the Hanson Lake pegmatite field; *in* Summary of-Investigations 1991, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 91-4.

Beryliferous granite pegmatite[2] dykes west of Hanson Lake (Figure 1) are confined to the quartz dioritic core of the Jackpine Lake Antiform. They strike north parallel to the axial plane of the antiform, dip exclusively east, and are traceable along strike for up to 500 m. The dykes tend to widen northwards to become the predominant bedrock in the northern part of the area. Asymmetrical textural and mineralogical zoning in the dykes is parallel to their sharp contacts, which locally crosscut the host-rock foliation. The pegmatites are composed of pink potassic feldspar, white albitic plagioclase, smokey to clear quartz, and pale green muscovite. Accessory biotite, garnet, magnetite and evenly distributed amounts of pale green euhedral beryl are also present (Pyke, 1966; Coleman *et al*. 1970; MacDougall, 1989).

Previous work showed that beryl mineralization is accompanied by significant elevated values of Li (max. 193 ppm), Nb (max. 142 ppm) and Sn (max. 222 ppm) (MacDougall 1989, Tables 8 and 9) and Ta (max. 216 ppm - this report, Table 1).

This year's investigations were aimed at determining the cause of the anomalous trace element geochemistry, completing the regional map and geochemical picture of the pegmatite field and finding out how the rare element concentration varies across and along the strike of the dykes. The work comprised (i) regional mapping (1:5000) and grab sampling; (ii) detailed prospecting, mapping (1:500) and channel sampling of two geochemically anomalous pegmatites (91-G and 91-H); (iii) detailed mapping (1:20) of a 20 m length of pegmatite 91-H.

1. Results of Mapping and Prospecting

a) Disposition of the Pegmatites

The pegmatite dykes are grouped together in braided zones up to 100 m wide, separated by barren areas 100 to 200 m across. Outcrop widths of individual dykes range from 5 cm to 12 m, though most are in the 1 to 4 m range. Narrow pegmatites are just as persistent along strike as the wider ones. Structural measurements (Figures 2 and 3) confirm that on a regional scale pegmatites and foliation are subparallel even though locally the pegmatites crosscut the foliation at high angles. Dip measurements on contact surfaces of pegmatite 91-H suggest that in general the western contact (40° to 72°) is steeper than the eastern one (<30° to 45°) and that the pegmatite widens at depth. The variation in dip along the contacts indicates billowing rather than planar pegmatite surfaces.

b) Internal Structure

Five distinctive lithologies are characteristic of the internal petrographic zoning of pegmatites throughout the Hanson Lake field:

1) Banded pink (or more rarely white), soda-aplite normally forms a broad (up to 1 m) persistent, but not entirely continuous zone along the western margin of the pegmatites. It is also repeated internally in places, and is present as scattered pockets along the eastern contact. The banding, which consists of single crystal layers of garnet or quartz-muscovite-garnet, tends to be concentrated near the eastern side of the zone (Plate 1; map sheet 2). Locally the aplite/granitic pegmatite contact is cusped and the banding cross—laminated-features which look practically sedimentary.

2) Granitic pegmatite forms a central zone in the pegmatites. It consists of a 1 to 2 cm quartz and potassic feldspar groundmass in which are set 10 to 30 cm (less commonly up to 60 cm), euhedral, pink potassic feldspars and sparse 5 to 10 cm muscovite books. Some of the feldspars are club-shaped, widen towards the centre of the pegmatite from their point of origin at the aplite/granitic pegmatite contact and show repeated chevron growth lines parallel to the crystal faces at their wide ends (Plate 1). Also present locally are euhedral quartz crystals, some of them showing concentric growth lines, and others which are intergrown with potassic feldspar and have a strange 'pagoda' shape.

3) Quartz, together with some potassic feldspar, forms a poorly developed, coarse-grained (5 to 30 cm), intermittent core zone within the granitic pegmatite.

4) Quartz-muscovite intergrowth commonly forms a discontinuous zone at the eastern margin of the pegmatites. It consists of a 1 to 2 cm groundmass of quartz and muscovite set with 10 to 30 cm euhedral to subhedral potassic feldspars. Some of the inter-

---

(1) Saskatchewan project A.148, from which this report derives, was funded in 1991 under the Canada-Saskatchewan Partnership Agreement on Mineral Development 1990-95.

(2) Throughout the text the term 'pegmatite' is used to mean granite pegmatite.

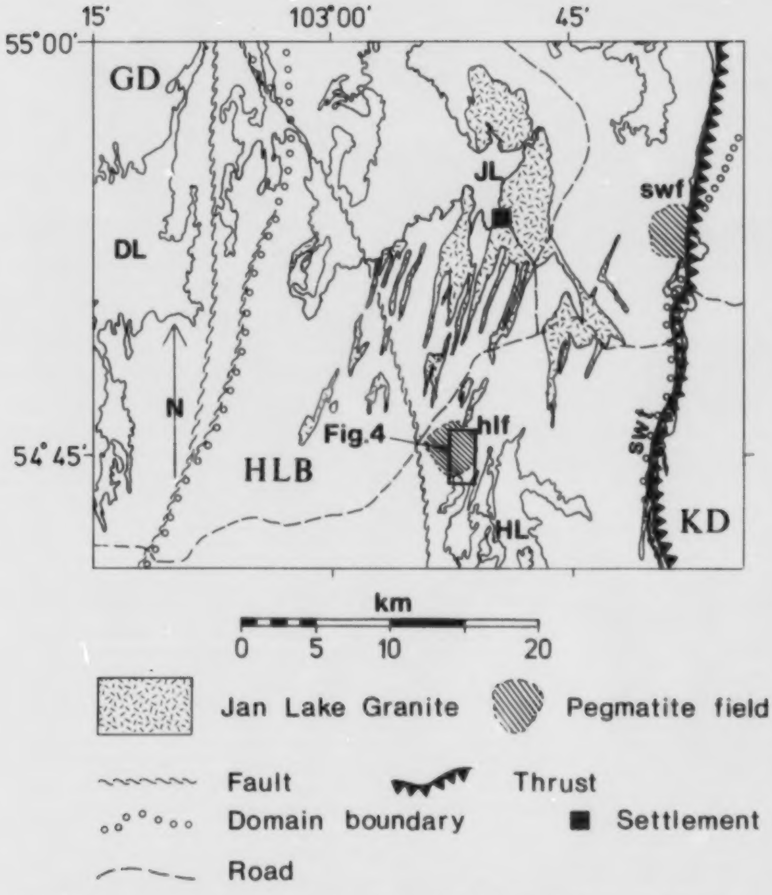


Figure 1 - Location of pegmatite fields in the Hanson Lake area. GD, Glennie Domain; HLB, Hanson Lake Block; KD, Kisseynew Domain: DL, Deschambault Lake; JL, Jan Lake and settlement of same name; HL, Hanson Lake; hlf, Hanson Lake pegmatite field; swf, Sturgeon-weir pegmatite field; swt, Sturgeon-weir Thrust.

growth is sheaf- or plume-like in form. The concentration of feldspar increases eastwards so that in places it forms a coarse mosaic against the eastern pegmatite contact.

5) Quartz-albite (var. cleavelandite) intergrowth, closely associated with quartz-muscovite intergrowth (Plate 2), becomes an important component of the pegmatites towards the northern part of the Hanson Lake field.

Small pockets of aplite, granitic pegmatite and white soda-aplite are present along the eastern contact of the pegmatites. Isolated chunks of banded aplite are present in the quartz-muscovite and central granitic pegmatite zones. Xenoliths of country-rock are absent. Small patches of quartz muscovite intergrowth occur in the granitic pegmatite zone.

Potassic feldspars in some of the narrow pegmatites (see MacDougall, 1989, Table 2) have a rubbly, crushed appearance, and have fine-grained albitic coatings. Some of the larger potassic feldspars in wide pegmatites are broken.

c) Mineralization

Beryl is present in similar concentrations in both narrow (5 to 30 cm) and wide (1 m and over) pegmatites. It is present in the granitic pegmatite zones, at the margins of quartz-rich core zones and as small patches of fine crystals in banded aplite. Beryl is typically pale green, inclusion-filled and opaque, but rare translucent to transparent beryl (one the variety aquamarine) is present in some quartz-rich core zones. Rarely, it is zoned and shows growth lines (Plate 3). As well as the more obvious scattered green beryls of 1 cm diameter and over, white beryl and numerous small

Table 1 - Hanson Lake Pegmatite Field Regional Trace Element Geochemistry (samples taken in 1989)

Sample No. ppm	89151	89155	89156	89158	89168	89171	89202	89208	89209	89213	89214	89220	89225	89233
Li	2	39	15	31	23	19	4	7	12	4	113	30	11	41
Nb	78	77	77	53	61	63	20	60	58	55	72	54	45	31
Ta	<1	27	170	<1	<1	50	<1	15	<1	<1	85	216	53	11
Sn	<20	53	48	75	<20	<20	<20	<20	29	<20	182	<20	21	<20
Sr	5	4	2	<1	27	<1	<1	6	<1	<1	<1	95	<1	<1
Rb	1078	1430	1239	671	1102	226	87	728	1602	164	1438	1368	2263	674
Cs	40	24	24	14	15	4	11	4	10	5	29	165	28	19

Li, Nb, Ta, Sn, Sr extraction by HF-HNO_3-$HClO_4$-HCl. Analytical method Induced Coupled Plasma - Atomic Emission Spectroscopy. Cs by Neutron Activation. Rb by X-ray Fluorescence. Analyses by Bondar-Clegg & Co.

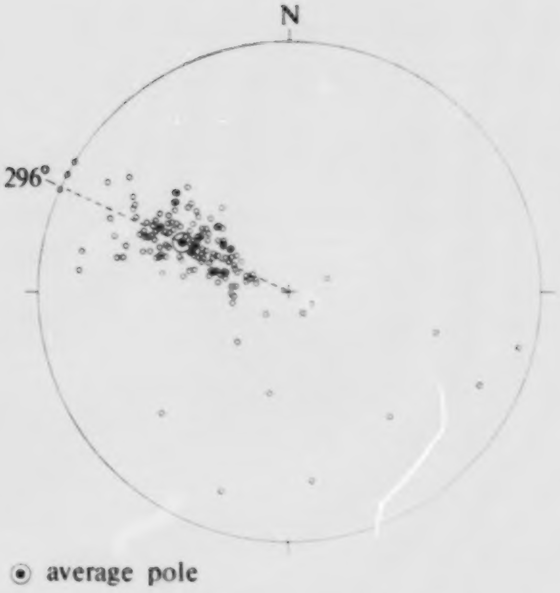


Figure 2 - Wulff Net stereographic plot of poles to 142 pegmatite contact surfaces. Over 90 percent of poles to pegmatite contact surfaces cluster around a pole to a plane striking 026° and dipping 50° southeast.

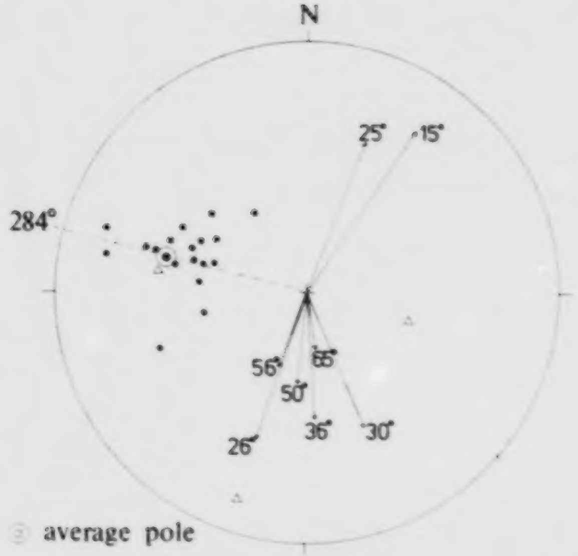


Figure 3 - Wulff Net stereographic plot of poles to 19 foliation planes (solid dots), 3 quartz veins (open triangles) and the axes of 8 minor folds (arrows). Poles to foliation cluster around a pole to a plane striking 014° and dipping 60° east-southeast. Minor fold axes fall into two groups, one northeast-plunging the other south-plunging.

(less than 5 mm diameter) pale inconspicuous beryls are also present. Geochemistry confirms this observation; significant beryllium values[3] are obtained from some samples containing no obvious beryl.

Recessively weathered, clay-altered, 1 to 3 cm tabular crystals of pale green spodumene were discovered in pegmatite 91-H (location 227 m N), associated with abundant white and green beryl, and up to 1 cm euhedral columbite. Euhedral and dendritic crystals of columbite 0.2 mm to 2 cm in size (Plate 4) are sparsely distributed in pegmatites throughout the Hanson Lake area.

Rare fracture-bound tourmaline (less than 1 mm euhedral grains), ubiquitous accessory gahnite spinel 0.5 mm in diameter and traces of euhedral fluorite 0.05 mm in size are apparent in thin section.

2. Geochemistry

Analyses of grab samples collected over the project area (Table 2) indicate Be values around 5.5 ppm which is consistent with the average for felsic igneous rocks (Hawkes and Webb, 1962). Three samples show significant beryllium enrichment (samples 91-0085, 35.8 ppm; 91-0091, 67.6 ppm and 91-0108, 141.8 ppm). The mean values of lithium (126 ppm), niobium (107 ppm) and tin (110 ppm) are respectively 2, 5 and 2.5 times the average for felsic igneous rocks (Hawkes and Webb, 1962), showing that there has been significant enrichment in these elements.

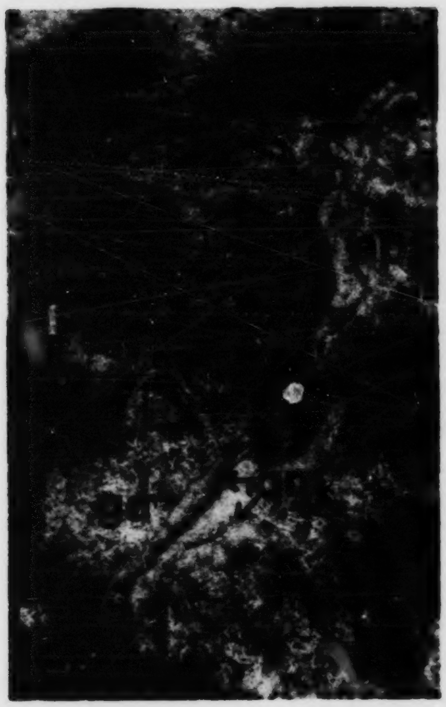
Analysis of muscovite from spodumene-bearing pegmatite (91-H, Table 2) shows that the mica is a rubidian variety, rich in lithium, tin and niobium. Muscovite can contain up to 5000 ppm tin and substantial niobium and tantalum as fine cassiterite grains containing columbite-tantalite inclusions (Ginzburg, 1954, 1955). The co-existence of spodumene and lithium-enriched mica shows that the available lithium was partitioned between silicate phases before the muscovite was saturated. According to Deer *et al.* (1966) muscovite can hold up to 3.5 percent Li_2O.

Exceptionally coarse grain size and penetrative weathering along feldspar cleavages and quartz-feldspar grain boundaries makes pegmatites difficult to sample effectively. Grab samples are seldom representative of the bulk mineralogy and chemistry. In an effort to overcome these problems, channel samples[4] were cut with a diamond saw across two of the more geochemically

---

(3) Beryl ($Be_3Al_2Si_6O_{18}$) has about 5 weight percent (50 000 ppm) beryllium. One cubic centimetre of beryl in a cube of rock (10 cm a side; the size of an average hand specimen) represents 1000 ppm beryl or 50 ppm beryllium. Beryllium does not fit in the lattices of any of the common rock-forming minerals but forms a common beryllium mineral, usually beryl. Analyses as low as 10 ppm beryllium can thus be taken to represent significant mineralization.

(4) Channels were 5 cm wide and about 5 cm deep; each 30 cm length along the channel was taken as a separate sample.



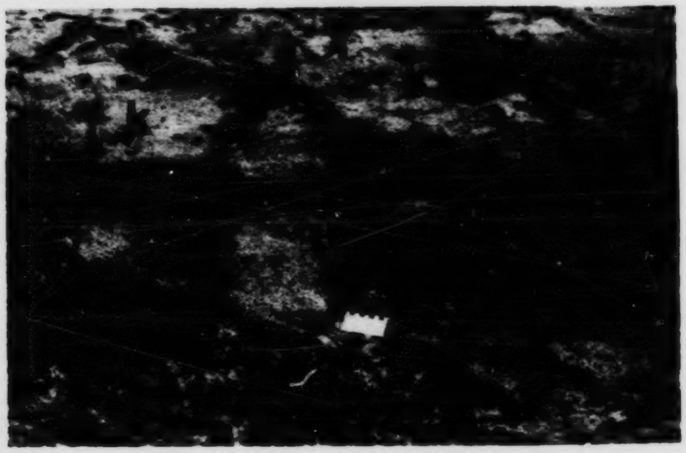


Plate 2 - Giant club-shaped potassic feldspars (crystal boundaries highlighted) in quartz-albite and quartz-muscovite groundmass pegmatite, northwest Hanson Lake area. Note that the feldspar on the left is broken k = potassic feldspars; qm = quartz-muscovite intergrowth; qa = quartz-albite intergrowth.

Plate 1 - Banded aplite and granitic pegmatite zones in pegmatite 91-H, 15 mN to 16 mN; gpz = granitic pegmatite zone; apz = banded aplite zone; k = potassic feldspars (crystal boundaries highlighted); ga = single crystal garnet band; qm = quartz-muscovite-garnet bands.

Plate 3 - Cluster of 2 to 3 cm diameter euhedral beryl crystals (be). Note the "carrot-core" concentric growth lines in the crystal to the right.

0 1 3 5 7
CENTIMETRES

Plate 4 - Dendritic and subhedral columbite in pegmatite, northwest Hanson Lake area. cb = columbite; se = sericite.

Table 2 - Hanson Lake Pegmatite Field Regional Trace Element Geochemistry (Samples taken in 1991).

Sample No. ppm	0084	0085	0087	0089	0090	0091	0092	0093	0094	0095	0096	0097	0099	0100
Be	5.4	35.8	5.5	6.4	5.1	67.6	7.4	5.5	7.0	3.9	6.4	3.8	9.4	7.3
Li	33	93	225	58	151	132	192	89	45	105	215	37	369	391
Nb	183	75	110	118	118	127	182	100	23	62	121	32	140	325
Ta	14	22	13	17	13	62	57	8	3	4	9	4	14	26
Sn	33	140	270	48	105	93	190	82	31	105	145	33	210	310
Sr	9	15	26	7	11	15	8	4	2	15	8	17	30	10
Rb	170	1450	1260	510	790	1840	670	270	72	1730	670	2010	1430	1230
Cs	3	17	15	5	7	21	8	2	7	11	6	12	16	11

Sample No. ppm	0103	0105	0108	0110	0111	0114	0116	0117	0118	0123	Muscovite 0082
Be	5.9	6.6	141.8	5.9	5.7	7.8	5.5	6.4	5.2	4.2	19.8
Li	42	245	44	5	2	14	36	347	110	54	1227
Nb	23	127	160	88	69	73	22	190	64	29	458
Ta	1	10	87	12	27	20	3	11	3	3	111
Sn	34	175	105	30	7	37	32	260	105	69	1000
Sr	9	32	67	14	14	28	10	5	13	4	39
Rb	110	640	3150	1170	1190	390	710	920	430	390	8200
Cs	1	6	54	15	11	6	7	8	5	5	174

Analytical methods; Be, Li by HF-HNO_3-$HClO_4$-HCl extraction and Atomic Absorption. Nb, Sn, Sr by X-ray Fluorescence. Ta, Rb, Cs by Neutron Activation. Analyses by Bondar-Clegg & Co.

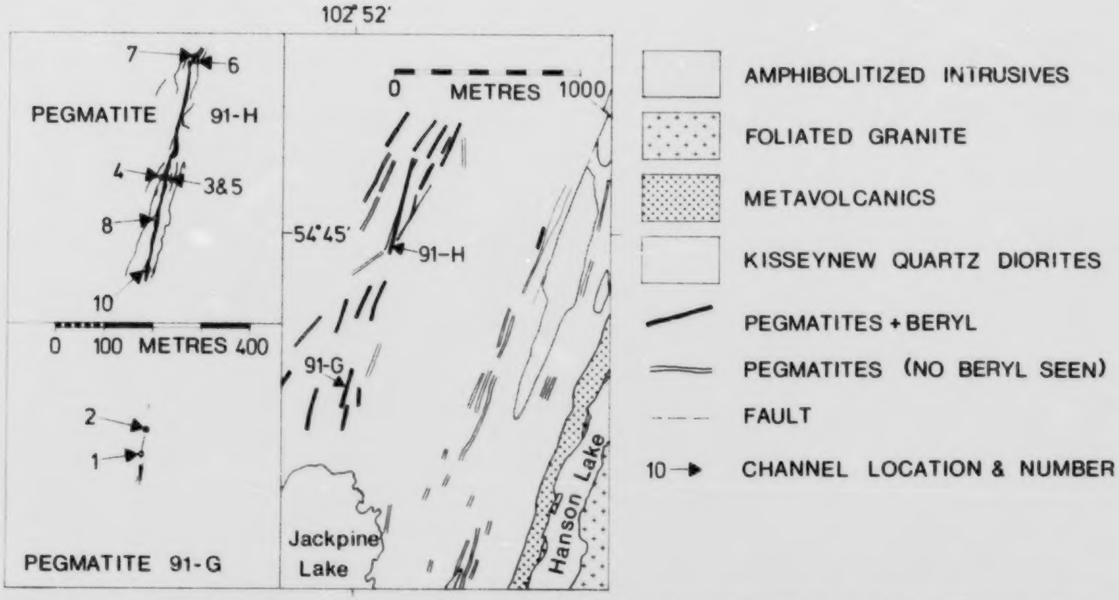

102°52'
7
6
PEGMATITE 91-H
4
3&5
8
10
0 100 METRES 400
2
1
PEGMATITE 91-G
0 METRES 1000
54°45'
91-H
91-G
Jackpine Lake
Hanson Lake
AMPHIBOLITIZED INTRUSIVES
FOLIATED GRANITE
METAVOLCANICS
KISSEYNEW QUARTZ DIORITES
PEGMATITES + BERYL
PEGMATITES (NO BERYL SEEN)
FAULT
10 CHANNEL LOCATION & NUMBER

Figure 4 - Map of part of the Hanson Lake pegmatite field showing location of pegmatites 91-G and 91-H. Insets showing location of geochemical sample channels 1 to 8 and 10.

anomalous pegmatites (91-G and 91-H; see Figure 4 and map sheets 1 and 2 for locations; see Figure 5 and Tables 3 and 4 for the results).

Li, Nb and Sn show a sympathetic relationship; values for all three reach a minor peak near the centre of the pegmatite and a maximum at the eastern margin. The close association of niobium and tin (as in the muscovite) may mean the mineral phase present is wodginite $(Ta,Nb,Sn,Mn,Fe)_{16}O_{32}$ rather than columbite-tantalite. Be and Ta show a sympathetic relationship, but not with Li, Nb and Sn. Be and Ta values reach a single pronounced peak to the west of the main Li-Nb-Sn peak, nearer the centre of the pegmatite. There is no variation in this pattern along strike.

Rb and Cs (Tables 3 and 4) vary sympathetically but, there is no systematic correlation with the other elements and the pattern differs for each channel. Rubidium and cesium values are either similar or sig-

Table 3 - Pegmatite 91-G Channel Sample Trace Element Geochemistry

	Channel 1	(90mN)	Channel 2 (181mN)	
Sample No. ppm	0013	0014	0015	0016
Be	69.5	297.3	18.8	203.8
Li	14	50	50	73
Nb	73	121	76	87
Ta	25	31	14	21
Sn	70	53	54	54
Sr	27	78	88	100
Rb	2190	650	1440	590
Cs	29	17	16	14

Compare Cerny *et al.*(1982) and Cerny and Meintzer (1988): Rb 32-1007 ppm, av. 346 ppm; Sr 2-115 ppm, av. 36 ppm; Cs 1.5-46 ppm, av. 9 ppm.

Analytical methods; Be, Li by HF-HNO_3-$HCLO_4$-HCl extraction and Atomic Absorption. Nb, Sn, Sr by X-ray Fluorescence. Ta, Rb, Cs by Neutron Activation. Analyses by Bondar-Clegg & Co.

nificantly higher and strontium values are significantly lower, than those reported by Cerny *et al.*(1982) and Cerny and Meintzer (1988) for a variety of rare element pegmatites and associated fertile granites. The high Rb and Cs and low Sr values indicate the source fluids responsible for the formation of the Hanson Lake pegmatites reached a high degree of differentiation.

3. Discussion

Any hypothesis for the formation of the pegmatites must explain the following features:

1) the mineralogical and textural zoning and its asymmetry;

2) concentric growth lines in beryl (Plate 3), quartz and potassic feldspar;

3) large size and euhedral shape of crystals;

4) absence of country-rock xenoliths;

5) discontinuity of banded aplitic zone, its repetition in the interior of the pegmatites, restriction to small pockets at the hanging wall contact, and presence as isolated chunks in the quartz-muscovite intergrowth and central granitic pegmatite zones;

6) the single crystal bands of quartz-muscovite-garnet and garnet in the aplite;

7) the cusped aplite/granitic pegmatite contact and the cross-laminated aplite banding;

8) the rubbly texture, crushed appearance and albitic coatings of the potassic feldspars in some narrow pegmatites, and the fragmentation of some of the larger feldspars in wide pegmatites;

9) quartz-muscovite intergrowth; and

10) the pattern of rare metal enrichment.

Radcliffe (1964) concluded that the pegmatites in the nearby Sturgeon-weir field grew inwards from the walls of open, fluid-filled fractures and that the beryl crystallized from large volumes of low-beryllium solutions moving through the fractures. The mineralogical and textural zoning, concentric growth lines in beryl, quartz and potassic feldspar, large size and euhedral shape of crystals and scarcity of country-rock xenoliths are all cited as evidence supporting crystallization from a low-viscosity, highly mobile fluid phase rather than a magma.

Certain features of the banded aplite; its discontinuity, repetition in the interior of the pegmatites and restriction to small pockets at the hanging wall contact, are inconsistent with a chilled-margin origin. Its precipitation was more likely triggered by critical changes in pressure, temperature and partial pressure of H_2O, the single crystal bands of quartz-muscovite-garnet and garnet in the aplite representing coatings deposited during repeated, but very temporary, special conditions.

The asymmetry of pegmatite zoning may have arisen because the open fractures were shallow-dipping rather than vertical. Material was precipitated and debris accumulated on the footwall "floor-zone" displacing the low viscosity pneumatolytic fluids towards the hanging wall "roof-zone", where their rapid movement tended to remove rather than deposit material. Restriction of aplite at the hanging wall contact of the pegmatites to small pockets and presence of isolated chunks of banded aplite in the quartz-muscovite intergrowth and central granitic pegmatite zones would be consistent with this process.

The rubbly texture, crushed appearance and albitic coatings of the potassic feldspars in some pegmatites and features which look like scour channels and cross-lamination in the aplite are consistent with brecciation, "fluid bed" movement and sedimentation by a fast-moving pneumatolytic fluid.

The quartz-muscovite intergrowth may represent the last phase of fracture filling by cotectic precipitation from a static solution, which after continued accretion of material on the fracture walls had choked off further influx of fluids. Where the volume of trapped fluid during this last stage, was sufficiently large, highly differentiated, rare metal-enriched zones developed.

4. Economic Potential

The economic potential of rare metal pegmatites is hard to assess. So few deposits are being mined that no established base line for comparison exists. The only mine in Canada producing rare metals (mainly tantalum but also on occasions lithium, cesium, rubidium, gallium and beryllium) from pegmatite is the Tanco Mine at Bernic Lake, Manitoba (Table 5). This pegmatite is a flatlying disc-shaped body, 450 to 1200 m in diameter and 18 to 85 m thick (Tantalum Mining Corporation of Canada Limited, publicity booklet). It lacks any apparent

Table 4 - Pegmatite 91-H Channel Sample Trace Element Geochemistry

Channel 3 a,b,c (three 30 cm channels at 224mN-226mN)				Channel 4 (west branch of 91-H at 226mN)			
Sample No. ppm	0028	0029	0030	0034	0035	0036	0037
Be	19.6	56.2	227.8	8.1	6.0	6.4	7.0
Li	106	777	394	113	37	67	158
Nb	74	174	83	89	61	41	117
Ta	31	85	109	17	11	7	21
Sn	32	140	60	75	37	62	81
Sr	27	21	33	85	10	15	129
Rb	3000	1920	3550	770	1310	2510	900
Cs	152	205	437	10	13	23	10

Channel 5 (east branch of 91-H at 226mN)											
Sample No. ppm	0038	0039	0040	0041	0042	0043	0044	0045	0046	0047	0048
Be	105.5	6.4	6.5	4.6	4.7	19.2	15.0	7.8	7.4	4.9	12.5
Li	46	69	194	134	63	105	56	283	297	73	81
Nb	124	94	240	103	109	153	114	140	211	60	87
Ta	106	17	36	18	22	34	48	20	23	8	22
Sn	29	18	56	44	29	64	27	300	90	65	74
Sr	44	13	3	10	18	5	7	16	16	36	70
Rb	2320	1720	290	1440	2430	730	720	2410	2080	3400	1590
Cs	159	48	12	25	65	14	10	32	22	25	15

Channel 6 (east branch of 91-H at 476mN)								
Sample No. ppm	0049	0050	0051	0052	0053	0054	0055	0060
Be	138.3	5.5	44.5	93.9	813.0	74.5	9.8	13.2
Li	43	38	30	30	62	37	177	1033
Nb	101	96	106	81	39	27	195	34
Ta	31	24	43	73	108	22	36	9
Sn	70	40	33	26	43	40	300	70
Sr	12	12	12	24	13	36	23	472
Rb	220	430	1010	2820	1470	4370	1330	1190
Cs	5	5	12	46	51	73	27	86

Channel 7 (west branch of 91-H at 483mN)					Channel 8 (interior of 91-H at 133.5mN)				
Sample No. ppm	0056	0057	0058	0059	0065	0061	0062	0063	0064
Be	42.0	14.1	6.4	96.2	6.5	5.6	4.4	230.4	6.6
Li	88	16	14	190	42	199	38	45	126
Nb	80	116	225	68	61	150	17	18	165
Ta	20	63	171	35	12	26	4	3	26
Sn	26	16	30	30	26	100	27	19	115
Sr	89	25	20	118	8	7	28	25	9
Rb	1420	1620	1600	1230	980	570	3690	4540	1360
Cs	13	13	12	20	8	8	31	78	12

Channel 10 (91-H at 19mN)										
Sample No. ppm	0068	0069	0070	0071	0072	0073	0074	0075	0076	0077
Be	303.6	116.0	22.5	13.2	10.0	140.0	1000	12.1	49.2	40.5
Li	56	91	75	148	28	182	148	175	305	283
Nb	115	78	111	133	72	117	133	134	161	34
Ta	112	20	30	28	21	59	99	26	28	10
Sn	18	51	48	96		165	175	120	190	36
Sr	66	17	9	6	1	12	10	8	14	181
Rb	2020	2130	1100	650	210	1630	1350	1170	1420	1720
Cs	37	25	25	15	17	47	36	12	14	22

Compare Cerny *et al.*(1982) and Cerny and Meintzer (1988): Rb 32-1007 ppm, av. 346 ppm; Sr 2-115 ppm, av. 36 ppm; Cs 1.5-46 ppm, av. 9 ppm.

Analytical methods; Be, Li by HF-HNO_3-$HClO_4$-HCl extraction and Atomic Absorption. Nb, Sn, Sr by X-ray Fluorescence. Ta, Rb, Cs by Neutron Activation. Analyses by Bondar-Clegg & Co.

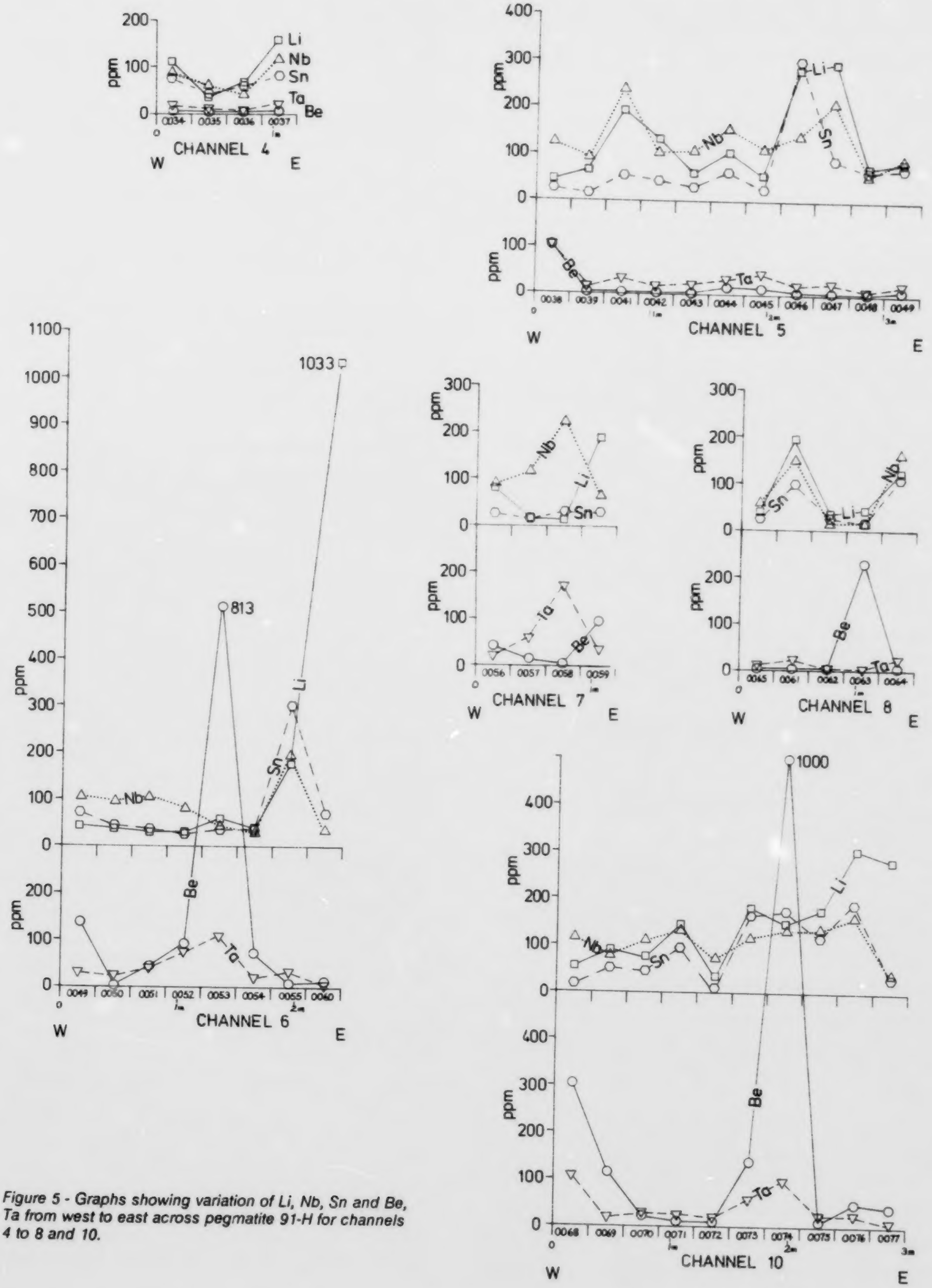


Figure 5 - Graphs showing variation of Li, Nb, Sn and Be, Ta from west to east across pegmatite 91-H for channels 4 to 8 and 10.

Table 5 - Ore and Mineral Reserves at the Tanco Mine, Bernic Lake, Manitoba (from Crouse et al, 1979)

Metal or Mineral	Tons	Grade	
Tantalum	2,071,358	0.216%	Ta_2O_5
Lithium	7,301,735	2.76%	Li_2O
Cesium	350,000	23.3%	Cs_2O
Beryllium	920,000	0.20%	BeO
Lepidolite	107,700	2.24%	Li_2O

feeder. Cut-off grades were originally calculated at 0.08 percent Ta_2O_5 (655 ppm Ta) and 1.00 percent Li_2O (4644 ppm Li) (Crouse *et al.*, 1979).

Other comparable pegmatite-hosted rare metal mines and deposits are:

1) The Wodgina Mine near Port Hedland, Western Australia which has reserves of 297 709 tonnes grading 0.206 percent Ta_2O_5. The nearby Mount Cassiterite and Tabba Tabba deposits have reserves of 1 million tonnes 0.08 percent Ta_2O_5 and 96 000 tonnes of 0.20 percent Ta_2O_5 respectively (Tolley, 1991).

2) The lithium pegmatites in the Kings Mountain area of North Carolina which contain 20 percent spodumene (0.6 percent Li) equivalent to 308 400 tonnes of lithium, 80 percent of U.S. reserves (Ferrell, 1985).

Pegmatites are not a primary source of beryllium but small quantities of beryl are produced by hand cobbing as a byproduct from pegmatites mined for their lithium or tantalum content (Petkof, 1985).

The potential for gem-quality beryl in the Hanson Lake pegmatites should be considered; gem-quality beryl has recently been discovered in pegmatites in Finland (Mining Journal, 1989) and has been known in Pakistani pegmatites since the 1960s (Bowersox and Anwar, 1989; Snee, 1991).

5. References

Bowersox, G. W. and Anwar, J. (1989): The Gujar Killi emerald deposit, Northwest Frontier Province, Pakistan; Gems & Gemology, Spring 1989, p16-24.

Cerny, P. and Meintzer, R. E. (1988): Fertile granites in the Archean and Proterozoic fields of rare-element pegmatites: crustal environment geochemistry and petrogenic relationships; *in* Taylor, R.P. and Strong, D.F. (eds.), Recent Advances in the Geology of Granite-Related Mineral Deposits, CIM, Spec. Publ. 39, p170-206.

Cerny, P., Roberts, W. L., Redden, J., Simmons, W. B., Foord, E. E. and Murphy, J. (1982): Granitic pegmatites of the Black Hills, South Dakota and Front Range, Colorado; Geol. Assoc. Can./Miner. Assoc. Can., Joint Annual Meeting, May 1982, Field Trip 12.

Coleman, L. C., Gaskarth, J. W. and Smith, J. R. (1970): Geology and geochemistry of the Hanson Lake area, Saskatchewan; Sask. Res. Counc., Rep. 10, 156p.

Crouse, R. A., Cerny, P., Trueman, D. L. and Burt, R. O. (1979): The Tanco pegmatite, southeastern Manitoba; CIM Bull., v72, No. 802, p142-151.

Deer, W. A., Howie, R. A. and Zussman, J. (1966): An introduction to the rock forming minerals; Longmans Publishers, 333p.

Ferrell, J. E. (1985): Lithium; *in* Mineral Facts and Problems, 1985; U.S. Dep. Inter., Bur. Mines Bull. 675, p461-470.

Ginzburg, A. I. (1954): About minerals-geochemical indicators and their significance in prospecting for ores of rare metals in pegmatites; Doklady Akademii Nauk SSSR, v98, No.2, p233-235.

__________ (1955): Summary of "About minerals-geochemical indicators and their significance in prospecting for ores of rare metals in pegmatites" (Ginzburg, 1954); Econ. Geol., v50, p748-750.

Hawkes, H. E. and Webb J. S. (1962): Geochemistry in Mineral Exploration; Harper and Row Publishers, 415p.

MacDougall, D. G. (1989): Beryl pegmatites in the Jan Lake area, NTS 63L and 63M; *in* Summary of Investigations 1989; Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 89-4, p71-81.

Mining Journal (1989): Finnish Emerald; Min. J., Dec. 1, v313, No. 8048, p440.

Petkof, B. (1985): Beryllium; *in* Mineral Facts and Problems, 1985, U.S. Dep. Inter., Bur. Mines Bull. 675, p75-82.

Pyke, M. W. (1966): The geology of the Pelican Narrows and Birch Portage areas, Saskatchewan; Sask. Dep. Miner. Resour., Rep. 93, 68p.

Radcliffe, D. (1964): The geology of the Birch Portage beryl pegmatite deposit, Saskatchewan; unpubl. M.Sc. Thesis, U. Alberta, 106p.

Snee, L. W. (1991): Emeralds of Pakistan and Afghanistan; Soc. Econ. Geol. Newsletter, No.5, April 1991, p20-22.

Tantalum Mining Corporation of Canada Limited, Bernic Lake Manitoba, undated publicity booklet.

Tolley, R. J. (1991): Specialty metals-tantalum; *in* Metals and Minerals Annual Review, 1991, p93-94.

Mineralized Core Collection

A.J. Gracie

Gracie A.J. (1991): Mineralized Core Collection; *in* Summary of Investigations 1991, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 91-4.

The Mineralized Core Collection has been described previously (Gracie, 1984, 1985, 1986, 1987, 1988, 1989, 1990) and an index to the La Ronge Core Collection also published (Gracie and Scutt, 1990). The index identifies the location of all core by longitude and latitude and on 1:250 000 scale maps, and gives full particulars of each hole: company, mineral dispositions, exploration commodity, year, hole number, core size, hole bearing and inclination, length of sample, reference to other exploration data, and storage location.

The mineralized core is housed in the Precambrian Geological Laboratory in La Ronge. Contributions to the Mineralized Core Collection totalled some 1973.2 m in 1990 (Table 1) and were made up of cores from mineral properties drill tested for gold (64 percent), uranium (25 percent) and base metals (11 percent). The Mineralized Core Collection is augmented by an extensive collection of rock samples from mineral properties in northern Saskatchewan. Important stratigraphic cores are housed in the Subsurface Geological Laboratory in Regina. No additions were made to this part of the collection last year.

Mineralized uranium cores were exhibited to a delegation nominated by the Ministry of Geology and Mineral Resources of the People's Republic of China and mineralized gold and base metal cores to participants on Field Trip 1 La Ronge Goldbelt and Field Trip 2 Flin Flon/Hanson Lake Area Massive Sulphide Deposits organized as part of the Geological Society of CIM First Annual Field Conference in Saskatoon. Also, a variety of mineralized cores were examined and sampled by research and exploration geologists.

1. References

Gracie, A.J. (1984): Mineralized Core Collection; *in* Summary of Investigations 1984, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 84-4, p127-128.

__________(1985): Mineralized Core Collection; *in* Summary of Investigations 1985, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 85-4, p82.

__________(1986): Mineralized Core Collection; *in* Summary of Investigations 1986, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 86-4, p121-122.

__________(1987): Mineralized Core Collection; *in* Summary of Investigations 1987, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 87-4, p80-81.

__________(1988): Mineralized Core Collection; *in* Summary of Investigations 1988, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 88-4, p125-126.

__________(1989): Mineralized Core Collection; *in* Summary of Investigations 1989, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 89-4, p110-111.

__________(1990): Mineralized Core Collection; *in* Summary of Investigations 1990, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 90-4, p118-119.

Gracie, A.J. and Scutt A.G. (1990): Index to La Ronge Core Collection; Sask. Energy and Mines, Data File 15, 228p.

Table 1 - Mineralized Core Acquisitions (September 1, 1990 to August 31, 1991).

Company or Joint Venture Operator	Commodity	NTS Area	No. of Holes	Core Size	Property/Deposit Location by Lake	Length (m)
Cameco	Au	73-P-16	3	NQ	Jasper Mine	291.6
	Au	73-P-16	2	NQ, BQ	Star Lake Mine	312.7
	U	74-H-10	1	NQ	McArthur River	200.7
	U	74-I-01	1	NQ	Waterbury Lake	138.7
Gamsan Resources	Cu, Zn	64-D-04	3	NQ	"McKenzie" Orebody	225.7
Gulf Minerals	U	64-L-05	3	BQ	Collins Bay	149.3
International Mahogany Minerals	Au	74-A-01	3	NQ	Jolu Mine	192.6
Placer Dome	Au	63-M-12	1	NQ	Seabee Mine	75.6
Radcliffe Resources	Au	73-P-07	3	IAX	"North" Lake	386.3

B. Northern Projects Under the Canada-Saskatchewan Partnership Agreement on Mineral Development 1990-95, from Year 1990-91

Geochemistry and Tectonic Setting of Metabasaltic Rocks from the Snake Rapids Area, Flin Flon Domain[1]

B.R. Watters[2] and K.E. Ashton

Watters, B.R. and Ashton, K.E. (1991): Geochemistry and tectonic setting of metabasaltic rocks from the Snake Rapids area, Flin Flon Domain; *in* Summary of Investigations 1991, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 91-4.

The Snake Rapids area is situated along the western side of Amisk Lake and represents the southwestern edge of the Flin Flon Domain. Supracrustal rocks in the area include: Amisk Group volcanic and volcaniclastic rocks in the east, a sedimentary sequence in the northwest, and a poorly exposed mixture of volcanic and sedimentary rocks south of the Sturgeon-weir River. These supracrustal units have been intruded by a number of large granodioritic and quartz monzodioritic plutons. For a more detailed description of the geology of the Snake Rapids area and the field relationships of the volcanic units, the reader is referred to Ashton (1990). The area has been included in geologic maps by Bruce (1918), Wright and Stockwell (1934), Byers and Dahlstrom (1954) and Fox (1976a).

Previous geochemical studies have indicated a dominance of pillowed mafic tholeiitic metavolcanics of probable island arc affinities (Fox, 1976b) in the southern part of the Snake Rapids area. A study by Walker and Watters (1982) in the adjacent area to the north provided evidence for an extensive sequence of dominantly intermediate to felsic volcanics of calc-alkaline character and volcanic arc affinity. It was suggested by Walker and Watters (1982) that the latter suite may represent the younger, upper and more emergent parts of an island arc, whereas the pillowed tholeiitic basalts could represent an older and predominantly subaqueous environment.

1. Present Study

The present geochemical study has been initiated to investigate the mafic metavolcanic rocks in the Snake Rapids area, recently remapped by Ashton (1990), in order to more clearly determine their tectonic setting and genetic relationships with the adjacent calc-alkaline volcanic suite.

For the purposes of geochemical analysis, samples were selected from the following locations: (a) the core of the Wolf Lake Fold where the best exposures of mafic volcanic rocks occur and where they are associated with minor felsic to intermediate volcanic units, (b) mafic units within the dominantly intermediate to felsic volcanic succession (calc-alkaline) in the eastern part of the area and representing the southward extension of the sequence studied by Walker and Watters (1982), and (c) the poorly exposed zone of volcanic rocks southwest of the Sturgeon-weir River.

Although the majority of samples analysed represent mafic metavolcanics, some represent associated intermediate and felsic metavolcanics. The latter were included to make comparisons with metavolcanic rocks from the dominantly intermediate to felsic succession to the north (Walker and Watters, 1982). Since the focus of the present study was to be those units comprising a dominance of mafic metavolcanics, no samples have been analysed from the intermediate to felsic metavolcanics in the northeastern part of the area (Ashton, 1990) which represents the southward extension of the area investigated by Walker and Watters (1982).

An initial suite of 38 samples have been analysed for major elements and the trace elements Rb, Sr, Y, Nb, Zr, Cu, Ni, Zn, Ba, V and Cr by XRF techniques at the University of Regina. Interpretation of the geochemical data is incomplete and only some preliminary results are reported here. Further analytical work, involving the determination of rare earth elements on a selected subset of samples, will be carried out in the near future in an effort to determine unequivocally the tectonic setting of the metavolcanic units.

2. Discussion of Results

Results obtained to date confirm a dominantly tholeiitic character for the volcanic rocks, although a minor calc-alkaline component is also present (Figure 1). For comparison purposes, data from the area immediately northeast of the Snake Rapids area, studied by Walker and Watters (1982), is also shown on the AFM plot (Figure 2) as well as a substantial amount of unpublished data. Although this diagram illustrates the dominantly calc-alkaline character of metavolcanic units in that area, the presence of a lesser component of associated tholeiitic mafic units is also notable.

A Jensen cation plot (Figure 3) for data from the Snake Rapids area shows results almost identical to the AFM plot. The similarity in the distribution of data points between the tholeiitic and calc-alkaline fields indicates that movement of the alkali elements in these rocks may not

---

(1) Saskatchewan Project B.107 of 1990-91, from which this report derives, was funded under the Canada-Saskatchewan Partnership Agreement on Mineral Development 1990-95.

(2) Department of Geology, University of Regina, Regina, Saskatchewan

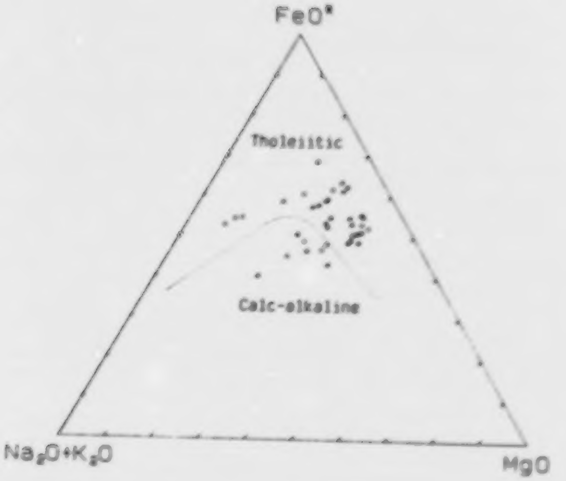


Figure 1 - AFM (Na_2O+K_2O-FeO^*-MgO) plot for metavolcanic rocks from the Snake Rapids area. Field boundaries are from Irvine and Barager (1971). Filled circles, samples from the dominantly mafic sequence in the core of the Wolf Lake Fold; open circles, mafic units interbedded within the dominantly intermediate-felsic sequence in the eastern part of the area; squares, metavolcanics southwest of the Sturgeon-weir River.

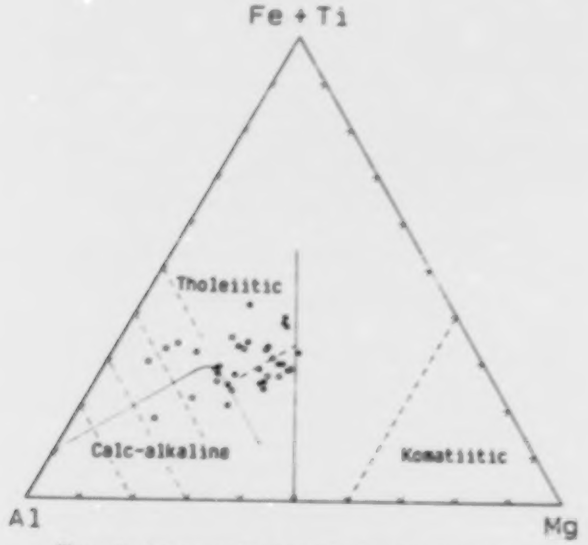


Figure 3 - Jensen cation plot for metavolcanic rocks from the Snake Rapids area. Field boundaries are from Jensen (1976). Symbols as for Figure 1.

have been substantial. On the basis of the samples analysed, it seems that the volcanic occurrence southwest of the Sturgeon-weir River may be composed entirely of tholeiitic volcanics, whereas those mafic (to intermediate) units in the core of the Wolf Lake Fold and to the immediate north (intercalated within the intermediate to felsic succession) comprise both tholeiitic and less abundant calc-alkaline units.

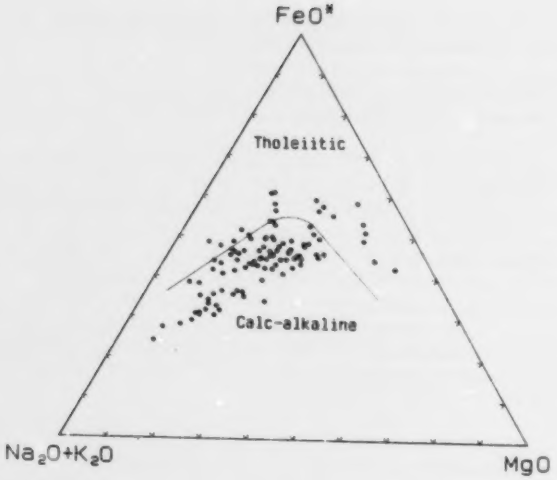


Figure 2 - AFM (Na_2O+K_2O-FeO^*-MgO) plot for metavolcanic rocks from the area immediately northeast of the Snake Rapids area (Walker and Watters, 1982). Field boundaries are from Irvine and Barager (1971). Symbols as for Figure 1.

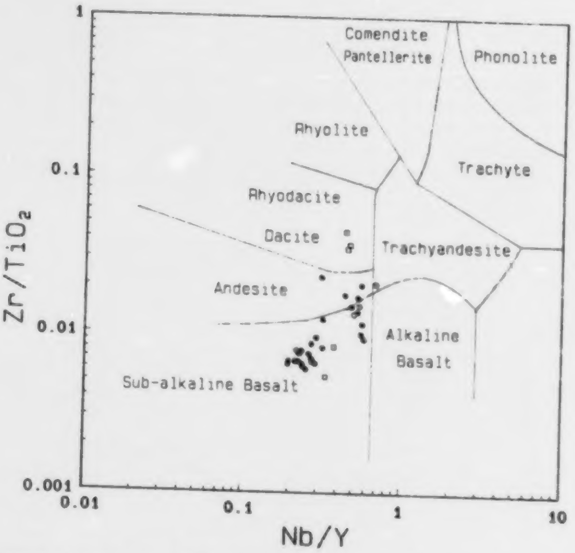


Figure 4 - Nb/Y vs. Zr/TiO_2 plot for metavolcanic rocks from the Snake Rapids area. Field boundaries are from Winchester and Floyd (1977). Symbols as for Figure 1.

The metavolcanics have relatively low Nb/Y ratios (Figure 4) and can be classified as sub-alkaline. On this diagram, which employs ratios between relatively immobile trace and minor elements, Nb/Y acts as an indicator of alkalinity. The ratio Zr/TiO_2 acts as a differentiation index and, in rocks which may have undergone alteration, is more reliable than SiO_2 which is commonly used for the same purpose. These trace and minor element relationships also indicate that the majority of the metavolcanic sequences are dominantly basaltic with minor andesitic and dacitic components.

The majority of the metavolcanic rocks are characterized by low to moderate Ti and Zr contents and fall mainly within the volcanic arc field and the overlapping MORB field (Figure 5). The more evolved rock types, including some of the basalts and all of the basaltic-andesites, andesites and dacites are clearly of volcanic arc affinity. Classification of many of the basaltic rocks is ambiguous however, in that they show Ti/Zr ratios that also characterize oceanic, rift-related basaltic rocks. Arc volcanic rocks typically have low TiO_2 contents (commonly less than about 1 percent) and, although this is the case for the majority of the Snake Rapids rocks, a significant proportion of them have relatively high concentrations of this oxide (up to 2.08 percent).

The Zr-Ti-Y discriminant plot (Figure 6) provides a similarly ambiguous tectonic classification for the metabasaltic rocks in that they cluster in the field common to both volcanic arc basalts and MORB. Concentrations of the element Y in the metabasaltic rocks are low to moderate with corresponding values of Cr that also suggest affinities to both volcanic arc and MORB character (Figure 7).

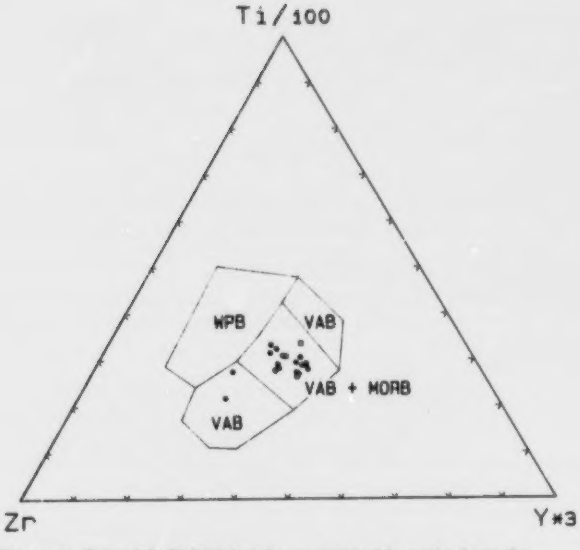


Figure 6 - Plot of Zr-Ti/100-Yx3 for metabasaltic rocks from the Snake Rapids area. Field boundaries are from Pearce and Cann (1973). VAB = volcanic arc basalt; MORB = mid-ocean ridge basalt; WPB = within-plate basalt. Symbols as for Figure 1.

MORB-normalized trace element patterns for selected samples of metabasalt (Figure 8) should be regarded as preliminary since many of the elements used in the array have not yet been determined. In spite of the poor representation of data, most of the profiles show a general pattern of mild enrichment in elements of low ionic potential (relatively mobile) and mild depletion in elements of high ionic potential (relatively immobile relative to typical tholeiitic MORB). Such patterns are similar to those shown by modern island arc tholeiites (Pearce, 1982); however, some obvious deviations can be noted: (a) Nb concentrations in the Snake River metabasalts are unusually high; typically, in volcanic arc basalts, Nb is less than unity when represented as a ratio to typical MORB values and; (b) some metabasalts show patterns with fairly strong enrichment in elements of low ionic potential and mild enrichment (except for Cr) in elements of high ionic potential. This type of pattern includes characteristics usually associated with within-plate-basalts and cannot be adequately accounted for at this stage except to speculate that they indicate a transitional volcanic arc environment.

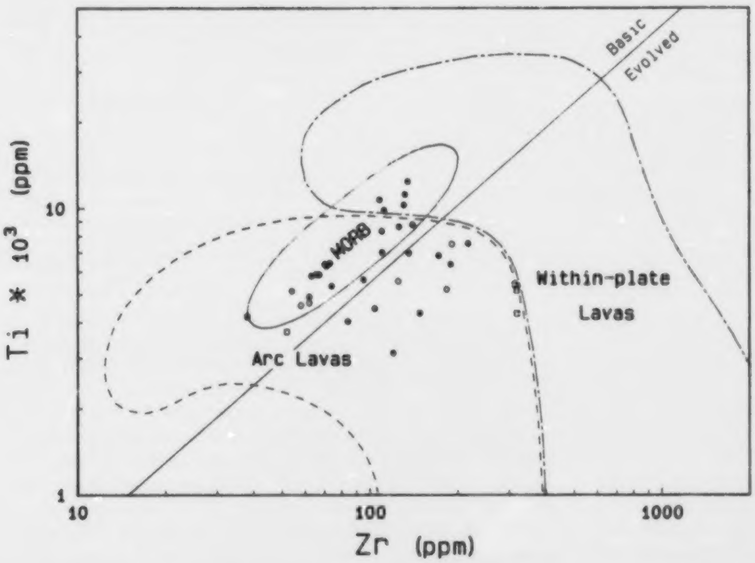


Figure 5 - Plot of Ti vs. Zr for metavolcanic rocks from the Snake Rapids area. Field boundaries are from Pearce and Norry (1979). Symbols as for Figure 1.

Metabasaltic rocks in the Snake Rapids area show geochemical characteristics that are similar in some respects to those of the East Amisk area (Gaskarth and Parslow, 1987). Those authors speculated on a possible back-arc environment having produced the transitional MORB-like characteristics. Such an interpretation may well be applicable to the Snake Rapids area, but the data available are more supportive of an island arc setting. Furthermore, the presence of calc-alkaline components within the dominantly tholeiitic succession, and the presence of tholeiitic flows in the dominantly calc-alkaline, intermediate to felsic succession to the immediate north, suggests a transitional relationship between the two sequences. Although such a relationship cannot entirely rule out the possibility of a back-arc setting (Pearce, 1982), it does suggest that a simple oceanic rift origin for the mafic rocks is unlikely. Structural data are not unequivocal, but do not contradict the suggestion (Walker and Watters, 1982) that the intermediate to felsic sequence may overlie the mafic (to intermediate) succession as would be expected in an evolving volcanic arc.

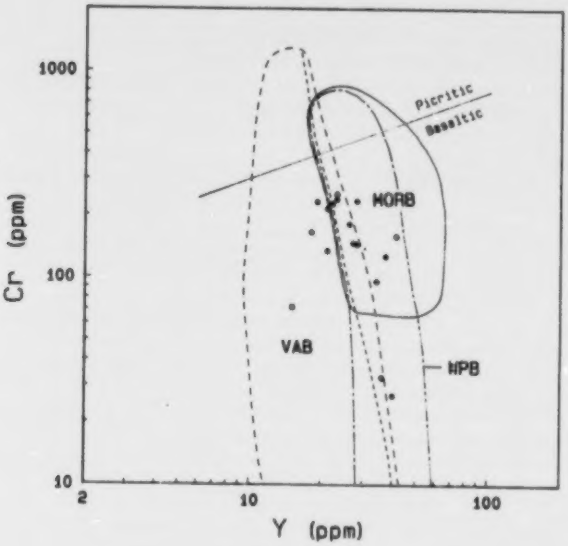


Figure 7 - Plot of Cr vs. Y for metabasalt rocks from the Snake Rapids area. Field boundaries are from Pearce (1982). Abbreviations as for Figure 6 and symbols as for Figure 1.

3. Conclusions

1. Metabasaltic rock units in the Snake Rapids area are tholeiitic and associated with lesser, calc-alkaline intermediate to felsic flows, which are geochemically similar to the bulk of the volcanic sequence immediately north of the study area.
2. It is tentatively suggested that the mafic metavolcanic rocks represent island arc tholeiites comprising the early, subaqueous stage of a developing volcanic arc. The dominantly intermediate to felsic succession to the north represents the later, mainly emergent stage. The data are not unequivocal; however, and a transitional environment or back-arc environment for the mafic units cannot yet be ruled out.

4. References

Ashton, K.E. (1990): Geology of the Snake Rapids area, Flin Flon Domain (parts of NTS 63L-9 and -10); *in* Summary of Investigations 1990; Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 90-4, p4-12.

Bruce, E.L. (1918): Amisk-Athapapuskow Lake district; Geol. Surv. Can., Mem. 105, 91p.

Byers, A.R. and Dahlstrom, C.D.A. (1954): Geology and mineral deposits of the Amisk-Wildnest Lakes area, Saskatchewan; Sask. Dep. Miner. Resour., Rep. 14, 177p.

Fox, J.S. (1976a): Preliminary bedrock geology of west Amisk Lake area, 63L-9 (W) and 63L-16 (SW); Sask. Res. Counc., Geol. Div., Misc. Rep.

________ (1976b): Volcanic stratigraphy and mineralization in the Amisk Group; Geol. Assoc. Can., Min. Assoc. Can., Joint Annual Meeting, Edmonton, Prog. Abstr., v1, p72.

Gaskarth, J.W. and Parslow, G.R. (1987): Proterozoic volcanism in the Flin Flon greenstone belt, east-central Saskatchewan, Canada; *in* Pharaoh, T.C., Beckinsale, R.D. and Rickard, D. (eds.), Geochemistry and Mineralization of Proterozoic Volcanic Suites, Geol. Soc., Spec. Publ. No. 33, p183-200.

Irvine, T.N. and Barager, W.R. (1971): A guide to the chemical classification of the common volcanic rocks; Can. J. Earth Sci., v8, p523-548.

Jensen, L.S. (1976): A new cation plot for classifying subalkalic volcanic rocks; Ont. Geol. Surv., Misc. Pap. 66.

Pearce, J.A. (1982): Trace element characteristics of lavas from destructive plate boundaries; *in* Thorpe, R.S. (ed.), Andesites, J. Wiley and Sons, New York, p525-547.

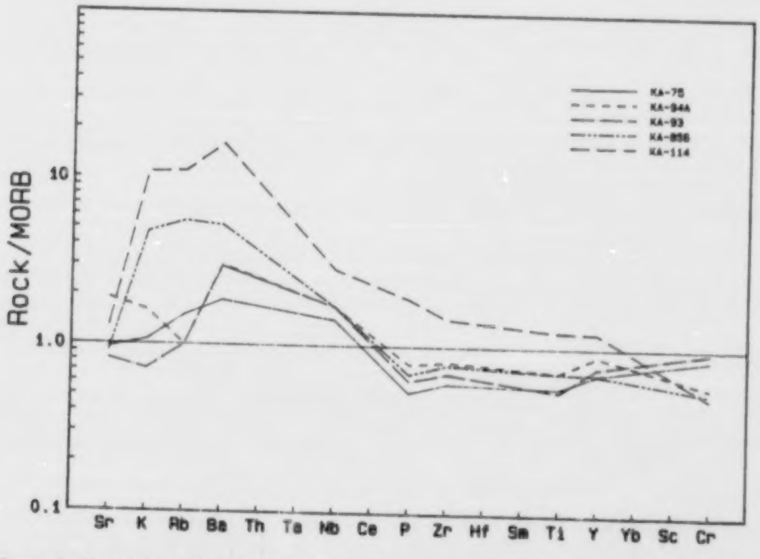


Figure 8 - Trace and minor element distribution patterns (MORB normalized) for some typical metabasalts from the Snake Rapids area. Normalizing values are from Pearce (1982).

Pearce, J.A. and Cann, J. R. (1973): Tectonic setting of basic volcanic rocks determined using trace element analysis; Earth Planet. Sci. Lett., vl9, p290-300.

Pearce, J.A. and Norry, M.A. (1979): Petrogenetic implications of Ti, Zr, Y and Nb variations in volcanic rocks; Contrib. Mineral. Petrol., v69, p33-47.

Walker, D. and Watters, B.R. (1982): Geochemistry of metavolcanic rocks, Amisk Lake west area; *in* Summary of Investigations 1982, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 82-4, p24-30.

Winchester, J.A. and Floyd, P.A. (1977): Geochemical discrimination of different magma series and their differentiation products using immobile elements; Chem. Geol., v20, p325-343.

Wright, J.F. and Stockwell, C.H. (1934): West half of Amisk Lake area, Saskatchewan; Geol. Surv. Can., Sum. Rep. 1933, Pt. C, p12-22.

Geological Relationships in the Wood Lake Area, Tabbernor Fault Zone, Saskatchewan[1]

K.H. Wilcox[2]

Wilcox, K.H. (1991): Geological relationships in the Wood Lake area, Tabbernor Fault Zone, Saskatchewan; *in* Summary of Investigations 1991, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 91-4.

During 1991 a 1:20 000 scale mapping project was continued in the region of the Tabbernor Fault Zone as part of a Ph.D. program at Carleton University. The purpose of the study is to investigate kinematic, metamorphic and geochronological features in part of the zone. The Tabbernor Fault Zone is a major north-south striking feature in Saskatchewan composed of lineaments, brittle faults, and plastic strain zones which are locally coincident with a rapid eastward increase in metamorphic grade. It has been used as a domain boundary separating the Glennie Domain on the west from the Kisseynew Domain and Hanson Lake Block on the east. General descriptions of rock units, kinematic features and metamorphic sequences observed during approximately four weeks mapping in the Wood Lake area in 1991 will be described here. The Tabbernor Fault described in this report refers to a major dominant fault within the Tabbernor Fault Zone which displays brittle features and generally a coincident negative topographic lineament.

Six main lithological subdivisions have been distinguished in the Wood Lake area (Figure 1). These are, in general chronological ascending order:

1) Prongua Lake Volcanic Belt
2) Tonalitic plutons
3) Wood Lake Granodiorite
4) Ourom Group
5) Transition rocks
6) Metasedimentary rocks

1. Prongua Lake Volcanic Belt (PLVB)

A belt of interpreted metavolcanogenic rocks and mafic sills which crops out from Prongua Lake to north of Wood Lake has been termed the Progua Lake Volcanic Belt. The belt is similar to the Lariviere Lake Volcanic Belt (LLVB), lying immediately west of the Tabbernor Fault south of the Wood Lake area. The PLVB is largely composed of discontinuous pods of plagioclase porphyritic mafic rocks considered to be plagioclase-phyric mafic intrusions, flows or sills (Wilcox, 1990). The shapes of plagioclase phenocrysts characterize variants that are generally only mappable on the outcrop scale. These variants contain phenocrysts which are fine-grained euhedral, medium-grained euhedral, medium-grained equant subhedral or coarsely orbicular. Fine-grained massive units are composed locally of interlocking plagioclase and hornblende. Epidotized pods and patches also occur.

Mafic tuffs to conglomerate layers form volcaniclastic horizons in the PLVB. Matrix-supported volcaniclastic conglomerates occur in beds up to 1 m thick. Oblate clasts include quartz diorite (dominant), epidosite, grey andesite and finely-layered turbidites. Quartz diorite clasts average 2.5 cm and as much as 45 cm in length. Sedimentary clasts are generally tabular. The matrix is composed of fine-grained hornblende, biotite and quartz. Conglomerate layers are commonly separated by tuffaceous beds containing angular equant plagioclase grains 1 to 2 mm in size which locally exhibit fining upward sequences.

Plagioclase-phyric and volcaniclastic rocks are cut by a number of pre-metamorphic intermediate to mafic dykes. A coarse-grained hornblende-bearing quartz diorite dyke with pyrrhotite concentrations and a similar metamorphic assemblage to the host metavolcanic rocks was observed on an island in Wood Lake.

2. Tonalitic Plutons

Medium-grained tonalitic plutons intrude volcanic rocks west of the Tabbernor Fault and metasedimentary units east of the fault (Wilcox, 1990). Although lithologically similar, it is uncertain whether intrusions on either side of the fault are of the same age. Further information on the tonalitic plutons is available in Sun *et al.* (this volume).

3. Wood Lake Granodiorite (WLG)

West of the Tabbernor Fault, the Wood Lake Granodiorite intrudes volcanic and volcaniclastic units of the PLVB. The granodiorite has yielded a U-Pb date of 1850 +1/-2 Ma in the Ourom Lake area (Delaney *et al.*, 1988). Kirkland (1976) mapped crosscutting lamprophyre dykes and gabbroic pods within the pluton.

---

[1] This project is supported in 1991-92 by the federal component of the Canada-Saskatchewan Partnership Agreement on Mineral Development 1990-95.

[2] Department of Earth Sciences, Ottawa-Carleton Geoscience Centre, Carleton University, Ottawa, Ontario, K1S 5B6

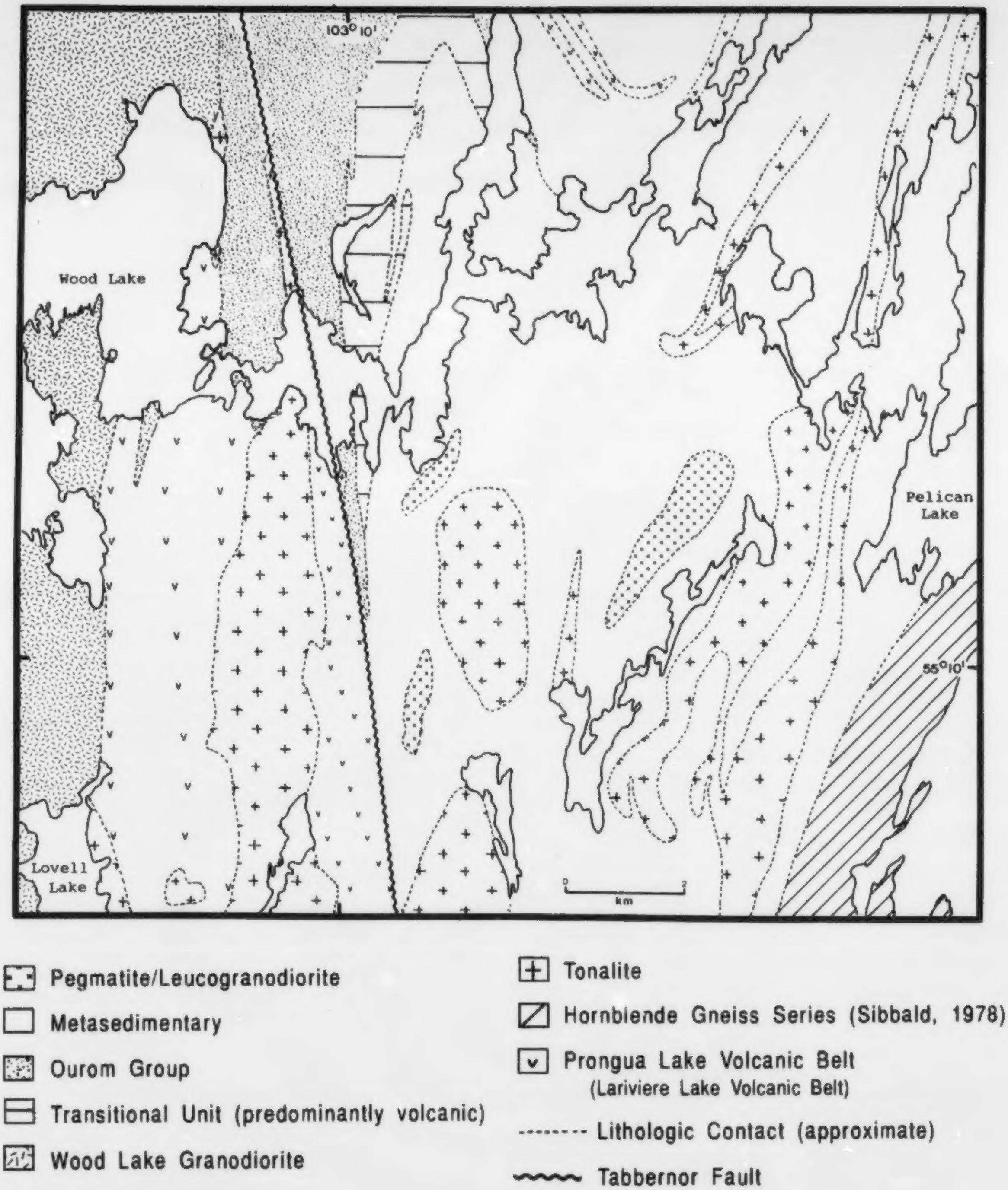

103° 10'
Wood Lake
Pelican Lake
55°10'
Lovell Lake
0
2
km
Pegmatite/Leucogranodiorite
Metasedimentary
Ourom Group
Transitional Unit (predominantly volcanic)
Wood Lake Granodiorite
Tonalite
Hornblende Gneiss Series (Sibbald, 1978)
Prongua Lake Volcanic Belt (Lariviere Lake Volcanic Belt)
Lithologic Contact (approximate)
Tabbernor Fault

Figure 1 - Generalized geology of the Wood Lake area. Geology from Kirkland (1976), Sibbald (1978), Wilcox (1990) and current work.

The Wood Lake Granodiorite is remarkably homogeneous and composed predominantly of subhedral interlocking quartz 1 to 4 mm in size, plagioclase and alkali feldspar with minor muscovite, rare biotite and accessory magnetite. Alkali feldspar grains exhibit the greatest range in size (up to 10 mm) and plagioclase is locally epidotized. Orange aplitic dykes occur locally. The main part of the pluton is not foliated but is locally well-jointed, and contains centimetre-scale fault zones and quartz veins.

Along the eastern margin of the pluton numerous dykes and sills of the granodiorite intrude metavolcanic and metavolcaniclastic units of the PLVB. The granodiorite dykes are slightly finer grained than the main body, in places contain a weak biotite foliation subparallel to the regional foliation, and are locally boudinaged. These dykes are crosscut by a rare generation of mafic dykes of unknown affiliation which crop out as greenish-black bodies composed of fine-grained interlocking hornblende and plagioclase.

4. Ourom Group

A sequence, mainly of meta-arkose and associated metasediments, crops out in an elongate north-south pattern between south of the Wood Lake area and northeast of Tabbernor Lake (Budding and Kirkland, 1956; Kirkland, 1957, 1976; Sibbald, 1978; Delaney, 1987; Wilcox, 1990). This belt is informally referred to as the Ourom Group. As the relationship of the Ourom Group to other units is integral to understanding the geological evolution of the area, a brief historical account is provided first.

Delaney (1987) proposed that the Ourom Group was deposited in an extensional basin having a western margin roughly coincident with the Tabbernor Fault Zone. Sibbald (1978) considered units in the Wood Lake area to be predominantly felsic volcaniclastics; he cited fragmental layers, possible fiamme textures, fine-grain size and lack of stratification as evidence. Budding and Kirkland (1956), Kirkland (1957), Sibbald (1978) and Delaney (1987) report that rocks associated with the Ourom Group are interlayered with volcanic and volcaniclastic units west of the Tabbernor Fault. South of Wood Lake a thin appendage of the Ourom Group parallels and is sinistrally offset by the Tabbernor Fault (Sibbald, 1978; Wilcox, 1990, this paper). The age of the Ourom Group is estimated to be ca. 1850 Ma by Delaney *et al.* (1988).

The author has recognized the following units in the Ourom Group in the Wood Lake area: 1) conglomerate, 2) grit to pebbly arenite, 3) interlayered mafic and felsic rocks, 4) arenite to arkose and 5) felsic plutonic or volcanic rocks.

a) Conglomerate

Three narrow and discontinuous conglomerate layers occur in the area: 1) at the contact with tonalite southeast of Wood Lake, 2) in contact with mafic volcanic rocks of the PLVB and 3) in a thin band within mafic volcanic rocks east of the Tabbernor Fault.

Conglomerate was observed as the basal member of the Ourom Group in unconformable contact with tonalitic rocks along the western shore of a peninsula immediately west of the Tabbernor Fault. At this isolated location the conglomerate is composed of rounded to elliptical clasts up to 15 cm long with approximately 3:1 aspect ratios. The unit is matrix-supported and contains intercalations of sandstone to mudstone. Clast types identified in the field include fine-grained polycrystalline quartz, coarse-grained polycrystalline quartz, a finely-laminated mudstone, mafic volcanic rock, fine-grained grey aphanitic material and possible white diorite.

A narrow layer of polymictic conglomerate less than [illegible]0 m thick was also observed at the contact with mafic volcanic rocks of the PLVB northeast of Wood Lake. Clasts set in a wacke matrix include fine-grained polycrystalline quartz, granitoid, fine-grained white diorite, plagioclase-porphyritic mafic volcanic rock and finely-laminated mudstone. Oblate clasts range from 1 to 20 cm in length and aspect ratios of up to 1:5 were observed in dioritic clasts. Isolated gritty layers up to 40 cm wide in the conglomerate are composed of granular 3 mm quartz.

b) Grit to Pebbly Arenite

Coarse sands to grit to pebbly arenite appear to be laterally equivalent to the conglomerate and form the dominant basal unit. This unit is more continuous than the conglomerate but still forms narrow discontinuous layers up to 100 m thick.

The matrix is composed of quartz 1 to 4 mm in size, and plagioclase grains with relict hematite and magnetite laminations which possibly represent trough to planar cross bedding. Beds from 5 to 50 cm thick are defined by abrupt and gradual grain size changes. Pebble horizons also occur locally. Isolated clasts include: coarse-grained polycrystalline quartz, fine-grained polycrystalline quartz, tonalitic rock, granodiorite, highly-flattened wispy black clasts of uncertain origin, grey volcanic rock (possibly andesite), laminated quartz (possibly rhyolite), jasper and oxide facies iron formation. Granitoid clasts tend to be largest and have the lowest aspect ratios. The quartzofeldspathic matrix of this unit has well defined muscovite partings.

c) Arenite-Arkose

The main unit is a fairly monotonous sequence of fine-grained arenite to arkose. Underlying coarser units appear transitional. Arenites and arkoses are commonly featureless, and composed of granoblastic plagioclase and quartz with rare hematite laminae presumed to be relict bedding horizons. Plagioclase grains are locally angular, imparting a tuffaceous or igneous appearance. When modal plagioclase exceeds quartz the unit commonly has a chalky weathered appearance. The unit contains less muscovite than basal units but has a well-developed parting.

d) Feldspar Porphyry

Small pods of felsic rocks near the base of the Ourom Group at Wood Lake locally contain quartz eyes and K-feldspar phenocrysts. Equant subhedral feldspar grains less than 3 mm long comprise less than 3 percent and quartz eyes 2 mm in size less than 2 percent of the rock. The phenocrysts are set in a salmon-coloured aphanitic matrix which has a muscovite parting. Weak layering defined by quartz-rich and feldspar-rich laminae is seen locally in the matrix. A lineation defined by quartz rods is also locally developed.

Three separate feldspar porphyry bodies were identified. Contacts were observed at one location immediately east of the Tabbernor Fault. The body at this location appears to be podiform, up to 20 m wide and 100 m long and surrounded by arenite or arkose. Contact relationships with the surrounding metasediments, are in general, obscured by high strain.

e) Felsic Intrusion

A quartz diorite to tonalite sill approximately 5 m wide and 50 m long crosscuts the Ourom Group north of Wood Lake, east of the Tabbernor Fault (UTM coordinates 6121830 mN, 614730 mE). It is composed of medium-grained quartz, plagioclase and hornblende. An orange-weathering aplitic phase also crosscuts tonalitic rocks.

f) Reserve Dykes

A mafic dyke suite crosscutting rocks of the Ourom Group, termed the Reserve Dykes, varies from massive, medium-grained hornblende-plagioclase rocks to an amphibole-chlorite schist. Rare relict feldspar phenocrysts up to 0.75 cm in length were observed locally. Post-kinematic garbenscheifer amphibole occurs in some of the schistose units. In regions of high strain adjacent to the Tabbernor Fault, dyke contacts are diffuse and chloritized. Incipient gneissosity defined by alternating amphibolitic and felsic layers also occurs in this area.

Budding and Kirkland (1956), Kirkland (1957, 1976) and Delaney (1987) report mafic dykes within the Ourom Group in other parts of the Tabbernor Fault Zone. Kirkland (1957) also reports a metamorphosed basic intrusion in the Ourom Group which is continuous into an amygdaloidal mafic volcanic rock. Kirkland (1976) interpreted abundant dykes and sills near the western contact of the Ourom Group to be the intrusive equivalents of basic metavolcanic rocks immediately to the west. A similar situation may exist in the Wood Lake area. Delaney (1987) considered that dioritic sills and dykes were intruded immediately after deposition of the Ourom Group and that the dykes may be evidence of a deep rooted fault system.

5. Transitional Unit

A highly-strained heterolithic unit dominated by well-cleaved intermediate to mafic rocks cropping out between the Ourom Group on the east and metasedimentary units on the west appears to be stratigraphically transitional with both. Included are mafic, felsic sedimentary and intrusive rocks. Intense deformation generally inhibits unequivocal interpretation of the protolith.

a) Mafic Rocks

Well-cleaved and locally well-lineated mafic rocks composed of amphibole, chlorite, plagioclase with or without biotite and carbonate dominate the eastern Transition rocks. Highly-flattened relict pillows were observed on two islands in Wood Lake, but in general primary features appear to have been eradicated during deformation and metamorphism. The rocks locally show elongate braided patterns defined by 2 cm wide plagioclase-carbonate networks separating amphibolitic pods. Massive sections composed of fine-grained amphibole and plagioclase were also found near the eastern margin of the unit. These rocks are interpreted as mafic volcanics.

Coarse hornblende-bearing gabbro with rare idioblastic garnet occurs as diffuse patches among mafic rocks north of Wood Lake. Contact relationships with the host mafic units were not observed.

b) Felsic Metasediments

A clastic unit composed of conglomerate and arenite is enclosed by well-cleaved mafic rocks immediately north of Wood Lake. The conglomerate is up to 10 m wide and contains clasts of oxide facies iron formation, granitoid, coarse-grained polycrystalline quartz, fine-grained polycrystalline quartz, fine-grained rock (possibly leucodiorite), fine-grained grey rock (possibly andesite) and arenite intraclasts. Laterally-associated arenite is dominant in the north. The arenite locally contains relict cross bedding defined by magnetite, but is generally massive with an appearance similar to strained leucocratic sills, from which it cannot in places be distinguished.

c) Garnet Tonalite

A distinctive and continuous sheet-like intrusive unit crops out from the southern shoreline of Wood Lake to the northern boundary of the study area. The rock is green on the weathered surface and composed of quartz, amphibole, plagioclase, chlorite biotite, and garnet. There is a well defined chlorite-amphibole foliation and strong quartz rod lineation. Equant plagioclase porphyroclasts occur locally. Idioblastic garnet crystals 1 to 3 mm in size are locally abundant and commonly concentrated in chlorite pods. A straight cleavage and weak gneissosity is developed in regions of elevated strain.

A pod of relatively low strained rocks is preserved along the north shore of Wood Lake. Down-lineation the rock shows relict igneous fabric defined by discontinuous networks of fine-grained polycrystalline quartz separating granular plagioclase, garnet and hornblende aggregates. Hornblende is locally absent and garnet aggregates are overgrown by chlorite. Garnet also occurs as isolated subidioblastic grains 3 mm in size; the relationship between garnet growth and the foliation is uncertain. The rock is dissected by braided high strain chlorite-biotite zones less than 15 cm wide.

d) Leucocratic Sills

Leucocratic sills, abundant near the garnet tonalite, and ranging from about 15 to 500 cm in width, are composed of granular quartz and plagioclase 1 mm in size.

6. Metasedimentary Rocks

Metasedimentary rocks lying east of the Tabbernor Fault and west of the Hanson Lake Block crop out as a southward-narrowing wedge contiguous with similar rocks of the Kisseynew Gneiss Domain (Macdonald and Broughton, 1980).

In the Wood Lake area, three subunits were recognized: 1) pelitic rocks with rare layers of greywacke, 2) subordinate interlayered pelite and greywacke and 3) coarse calc-arenite and pebbly conglomerate.

Although the contact between the metasedimentary rocks and the Mafic Rocks east of the Ourom Group coincides with an area of high strain, interlayering of mafic and pelitic units suggests that the contact is stratigraphic. Coarse clastic rocks occur as discontinuous layers in the pelites near the volcanic rocks and abruptly disappear eastward where the pelites are dominant.

a) Pelites

Pelitic rocks display an eastward increasing metamorphic grade. At lowest grade, which occurs immediately east of the Tabbernor Fault or locally along the eastern contact with metavolcanic units, a discontinuous phyllite to fine-grained schist crops out. Centimetre-scale bedding is locally evident and foliation locally crenulated. Matrix minerals immediately coarsen eastward to a medium-grained schist (quartz, plagioclase and biotite) where fine-scale bedding is progressively difficult to determine. Aluminous units contain andalusite porphyroblasts which are progressively pseudomorphed to the east by sillimanite and polycrystalline aggregates. Variations in bulk composition of the metapelitic units near the contact with the mafic rocks have produced local concentrations of hornblende, garnet, staurolite, cordierite, andalusite and sillimanite porphyroblasts. Post-kinematic anthophyllite transecting the foliation also occurs. Coarsely crenulated schists which locally contain cordierite, sillimanite and garnet are common throughout the study area to the east.

A finely-laminated hornblende-plagioclase layer, associated with hornblende-bearing biotite schists, occurs in the pelitic unit between the northeast arms of Wood Lake and was mapped as a volcaniclastic rock.

Calc-silicate assemblages are evident throughout the pelitic rocks, occurring as weakly elliptical to circular annular pods less than 15 cm in diameter. Individual minerals are unoriented within the pods and are commonly composed of monominerallic garnet, hornblende, a brown amphibole, plagioclase, epidote or diopside. The calc-silicate pods are interpreted as primary calcareous concretions.

b) Interlayered Pelite and Greywacke

Intercalated pelitic schists and greywacke occurring in layers less than 2 m wide form a mappable unit near the eastern edge of the map area. Pelite layers contain coarse quartz, plagioclase, biotite, faserkiesel, cordierite and garnet, and are in sharp contact with greywacke layers composed of granoblastic quartz 1 mm in size, plagioclase with disseminated biotite, and rare garnet.

c) Conglomerate and Calcarenite

Two separate coarse horizons were recognized in the metasedimentary sequence.

1) A small elongate pod of coarse sedimentary rocks is interlayered with pelitic rocks adjacent to the Transition rocks. Centimetre- to metre-wide layers include arenite, pelite, grit, pebbly wacke and conglomerate. Clast size in conglomerate horizons ranges from sand-sized to 10 cm long and averages 2 cm. Clast types include polycrystalline quartz, mud intraclasts, aphanitic felsic clasts and opalescent blue quartz granules. Mud clasts are typically tabular. Polycrystalline quartz clasts are less than 1 cm long. Blue quartz granules average 2.5 mm in diameter. Rare garnet and hornblende porphyroblasts occur in the pebbly wacke. Metapelitic rocks are mainly composed of muscovite and biotite.

2) A distinctive coarse unit in the centre of the map area is defined by interlayered beds of oligomictic pebble conglomerate, calcarenite and locally greywacke. Calcarenite layers are commonly massive, composed of granoblastic quartz, plagioclase, hornblende and biotite, and vary from 40 to 1000 cm in width. Conglomerate layers 15 to 100 cm wide lie in sharp contact with the calcarenite and locally exhibit upward-fining sequences. Upward-fining sequences at Wood Lake (coordinates 6119700 N 18200 E) are approximately 1 m wide. These have basal contacts with the underlying arenite and are overlain by pebble conglomerates 5 to 10 cm wide composed of approximately rounded clasts 5 mm long in point contact. The layers grade into slightly coarser matrix-supported pebble conglomerate containing clasts averaging 2.5 cm in size set in a grey-blue arenite to greywacke matrix. There is a 15 cm wide transition to arenite. Weakly developed tabular crossbedding is apparent at the top of the arenite and is rarely capped by a finely-laminated mudstone 4 cm thick. Rounded clasts in the conglomerate include polycrystalline quartz, aphanitic felsic material, fine-grained granitoid, gabbroic-textured hornblende-plagioclase, quartz diorite, and calc-silicate (equigranular intergrowths of plagioclase, hornblende and garnet). Coarse hornblende porphyroblasts up to 6 mm long and garnet within felsic clasts are restricted to pebbly horizons.

7. Structure

Wilcox (1990) interpreted structures in terms of at least two deformation phases - an initial compressional event responsible for tight to isoclinal folds of bedding, and a second long-duration and kinematically complex event related to the Tabbernor Strain Zone. The Tabbernor Strain Zone is defined as the area where heterogeneous plastic and brittle strain features were developed during the latter event. Plastic mesoscopic features occurring in rocks of higher metamorphic grade in eastern regions of the Tabbernor Strain Zone indicate reverse or oblique sinistral reverse displacement, whereas brittle-plastic sinistral strike-slip displacement is recorded around the Tabbernor Fault in regions of lower metamorphic grade. The Tabbernor Fault was formed during the terminal stages of the brittle deformation in the Tabbernor Strain Zone.

Similar kinematic and rheological features were observed in 1991. The zone exhibiting the highest plastic strain does not follow the Tabbernor Fault, but bifurcates to the northeast (Figure 2). With increase in metamorphic grade east of the zone, discrete plastic strain features die out. Discrete brittle features, strongest at the Tabbernor Fault, disappear to the west with decrease in metamorphic grade.

a) Tabbernor Strain Zone

Although the degree of strain in the Transition rocks between the Ourom Group and Metasedimentary rocks is difficult to assess, heterogeneously strained garnet tonalite indicates high strain in this area. Garnet tonalite contains penetrative quartz rods and hornblende and biotite streaking lineation which plunges 70° to the southeast. Kinematic indicators in this unit, found always in association with an extension lineation, include S-C fabrics (Berthe *et al.*, 1979), sinistral asymmetric extension cleavage (shear bands) (Platt and Vissers, 1980), asymmetric boudins (Hanmer, 1986), and rare asymmetric folds. These features consistently indicate sinistral reverse oblique shear along steeply east-dipping planes.

In the metasedimentary rocks to the east, a well developed lineation is defined locally by pebble and quartz rods, fibrolite, and elongate aggregates of biotite. This lineation is colinear with the hinge lines of tight folds, which are both less penetrative to the east where metamorphic grade is higher. The metasediments do not appear as highly strained as the adjacent Transition rocks, and do not generally contain mesoscopic shear sense indicators. Foliation is also penetratively crenulated about variable northeast- to northwest-striking hinge surfaces.

Lineation progressively shallows within the Ourom Group west of the body of garnet tonalite and is locally subhorizontal adjacent to the Tabbernor Fault. Although high strain zones and sense of shear indicators are relatively rare in the Ourom Group east of the Tabbernor Fault, rare asymmetric extension cleavage indicates a shear sense of the east side up and to the north.

Brittle-plastic strain zones in rocks adjacent to the Tabbernor Fault locally contain shallow-plunging mineral lineations. Mesoscopic strain zones 10 cm to 25 m wide locally contain S-C fabrics, asymmetric extension cleavage, asymmetric folds, foliation deflections, rotated cobbles and asymmetric boudins all of which indicate sinistral strike-slip sense of shear. Pebbly arenite to grit units west of the Tabbernor Fault at Wood Lake exhibit penetrative mesoscopic Z folds in bedding with steep hinges and axial planar cleavage defined by muscovite. This suggests that the strain was possibly pure shear and coaxial. Asymmetrically boudinaged clasts, fabric deflections along cleavage planes and sinistral asymmetric extension cleavage coeval to the folds, on the other hand, indicate sinistral strike-slip displacement. Thus the fold asymmetry in this region is probably a result of elevated non-coaxial shear strain (Ramsay *et al.*, 1983).

Tonalitic rocks on the west side of the Tabbernor Fault south of Wood Lake contain abundant centimetre-scale discrete strain zones in braided patterns. The zones have abrupt walls. Slickenside lineation and internal (S fabric) chlorite is oriented obliquely to these walls. Asymmetric extension cleavage and late stage epidotic alteration found in many zones respectively aid and inhibit kinematic interpretation. Lineations in NNE-striking zones commonly plunge at low angles to the north. In NNW-striking zones the lineation plunges subhorizontally. The oblique foliation indicates sinistral shear in NNW-striking zones and dextral shear in NNE-striking zones. Discrete zones are replaced by irregularly and discontinuously cleaved areas where the two sets intersect. The continuation of the same tonalite body north of Wood Lake is 50 m wide and 2 km long and contains penetrative asymmetric extension cleavage indicating sinistral strike-slip sense of shear. Brittle features and associated alteration are generally confined to the area around the Tabbernor Fault and include small-scale faults, breccia bodies, pseudotachylite, silicification, epidotization and hematite alteration. A brittle-plastic strain zone within the Ourom Group at Crooked Lake contains abundant silicification with secondary feldspathization, epidotization and hematization. Milky-white quartz veinlets impart a bleached appearance to the rock.

Discrete centimetre-wide NNW-striking fault zones are locally evident in WLG dykes and eastern regions of the unfoliated WLG. Chlorite oriented obliquely to the fault zone walls indicates apparent sinistral displacement in these zones. The zones are not present in the western part of the pluton.

8. Metamorphism

West of the Tabbernor Fault metamorphic grade appears to be fairly uniform at lower amphibolite facies in volcanogenic rocks of the PLVB. Metamorphic grade in the Ourom Group is low on both sides of the Tabbernor Fault.

a) Isograds

The garnet isograd comes in on Wood Lake in highly-strained Transition rocks lying east of the Tabbernor Fault and 250 m east of the Ourom Group. The garnet here is less than 4 mm in size and disseminated as subidioblastic grains. The isograd trends obliquely to the Tabbernor Fault.

Three isograds and key mineral textures occur in the Metasedimentary Units.

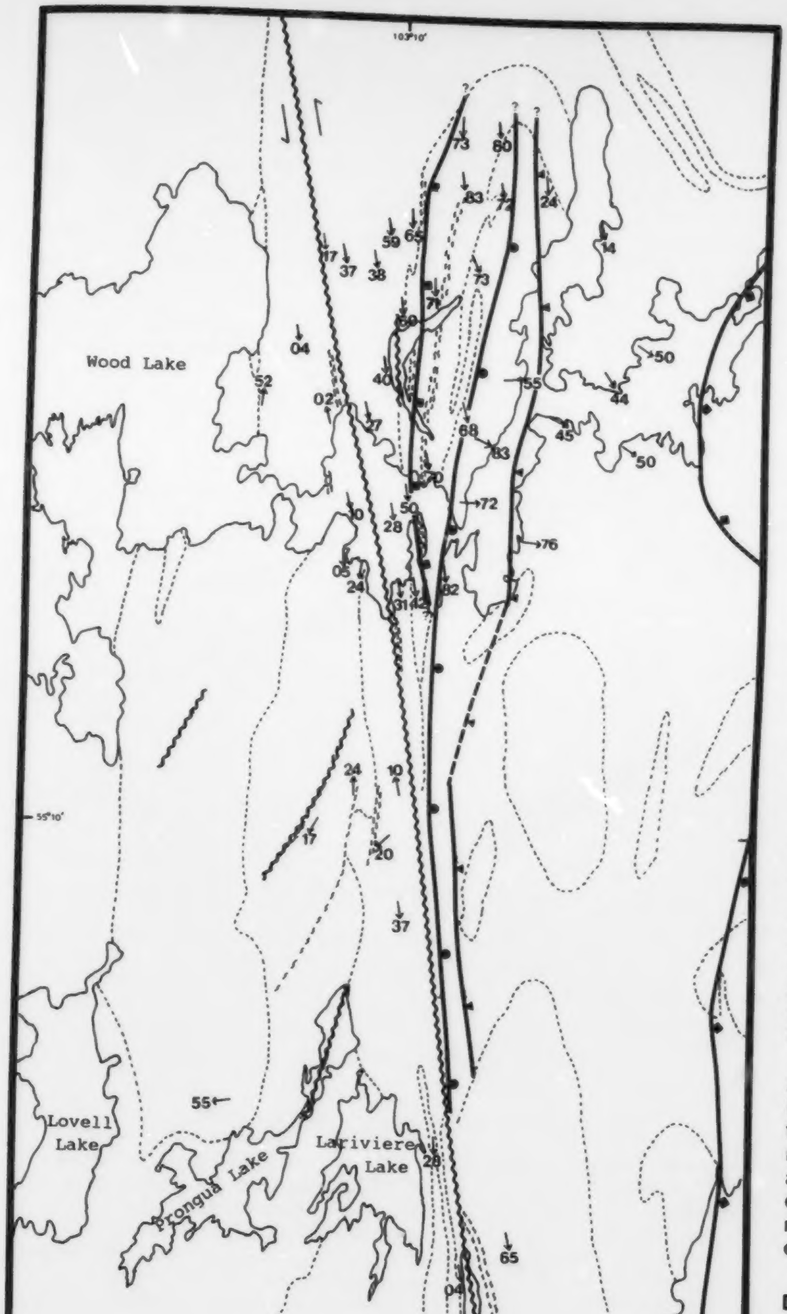

Legend
Predominantly Brittle Fault (unknown, sinistral)
Predominantly Plastic Strain Zone
Lithologic Contact
Extension Lineation
Garnet Isograd
Aluminosilicate Isograd
Andalusite "Out" Isograd (approximate, interpolated)
Cordierite Isograd
103°10'
55°10'
Wood Lake
Lovell Lake
Lariviere Lake
Prongua Lake
0
4
km

Figure 2 - Location of high strain zones and isograds in the study area. Geology from Kirkland (1976), Sibbald (1978) and Wilcox (1990 and current work).

Andalusite-Sillimanite Isograd

Andalusite and sillimanite occur together at the first appearance of an aluminosilicate east of the phyllite outcropping near the contact of the Metasedimentary Unit and the volcanic PLVB. The first occurrence of andalusite is as dark prismatic porphyroblasts approximately 1 cm long which are generally unoriented in a fine-grained matrix of uncertain mineral composition. Porphyroblasts coarsen to lengths of about 3 cm over a distance of 10 m. This feature is associated with coarsening of matrix quartz, plagioclase and biotite. Andalusite also occurs in these rocks as quartz-andalusite (chiastolite) pods and coarse pink chiastolite grains up to 6 cm long overgrowing early andalusite in the matrix. Quartz-andalusite pods are locally folded or foliation deflects around them.

Andalusite is most abundant and best preserved immediately east of the isograd and progressively replaced to the east. The aluminosilicate "in" isograd lies oblique to the Tabbernor Fault and represents at least two growth episodes in this area.

Disappearance of Andalusite

An isograd representing the complete disappearance of andalusite defines the eastern margin of a 1 km wide zone of partially preserved andalusite. The isograd is marked either by progressive pseudomorphing of andalusite by sillimanite or by polycrystalline aggregates (see Wilcox, 1990).

Rounded to elliptical faserkiesel also are abundant near the andalusite "out" isograd.

Cordierite Isograd

Cordierite has two distinct growth associations, both of which occur in a coarse crenulated biotite-quartz-plagioclase matrix.

One cordierite association is in quartz-plagioclase-muscovite segregation pods to veins which comprise less than 1 percent of the total rock. Cordierite ranging in size from 0.5 mm to 50 mm composes from less than 1 to 25 percent of the pods. Two generations of veins are recognized. An early vein set lies generally parallel to and folded with the foliation while second generation veins are commonly oriented axial planar to the folds.

The other occurrence of cordierite is as purple translucent grains partially or wholly replacing faserkiesel (Sibbald, 1978). The cordierite isograd occurs over 2 km east of the andalusite "out" isograd and appears to strike more northeasterly than other isograds.

b) Mineral Textures

Sporadic staurolite is restricted to ferruginous metapelitic rocks generally between the aluminosilicate "in" and andalusite "out" isograds. It is always rimmed by sillimanite and in various stages of breakdown. Similar to andalusite, staurolite is progressively consumed to the east and has not been observed where cordierite appears stable. Garnet is a common phase in these localities, occurring as idioblastic grains 2 mm in size which locally appear to interrupt the foliation.

In regions of cordierite stability, garnet is a common minor phase occurring as grains 8 mm in size invariably rimmed by sheaths of sillimanite (and possibly quartz) which rarely contain cordierite.

Muscovite occurs in all metasedimentary rocks, generally within 1 km of the Tabbernor Fault, as coarse flakes or books along or transecting cleavage planes.

Anthophyllite occurs as brown isolated to radiating blades, unoriented and transecting the foliation. It was observed in most metapelitic rocks between the aluminosilicate "in" and andalusite "out" isograds. Two anthophyllite-bearing assemblages occur with matrix hornblende in calcareous wackes and (in three localities) with cordierite . At cordierite-anthophyllite locations, cordierite occurs as large inclusion-riddled porphyroblasts locally overgrown by coarse anthophyllite up to 7 cm long. Cordierite-anthophyllite assemblages are commonly associated with metamorphosed magnesium alteration zones around massive sulphide deposits. Such assemblages have also been associated with regional metamorphism of Mg-rich metapelitic rocks (Reinhardt, 1987) and metasomatism during deformation (McCaig, 1987). The presence of late stage crystallization and disequilibrium (with hornblende) assemblages may indicate that anthophyllite was not produced as a result of initial bulk composition.

9. Summary

Preliminary assessment of the Wood Lake area indicates that an early volcanogenic assemblage, the Pongua Lake Volcanic Belt, was intruded by tonalitic plutons and later by the Wood Lake Granodiorite. The Ourom Group appears to have been unconformably deposited on tonalite and mafic volcanic rocks of the PLVB along its west margin, but was probably mainly derived from arenaceous detritus from the Wood Lake Batholith. The Ourom Group may also have been deposited contemporaneously with Transition rocks and portions of a Sedimentary Sequence. Interlayered and locally subaqueous, volcanic rocks and felsic sediments on the eastern margin of the Ourom Group suggests that this area was a site of mixed volcanism and sedimentation. A similar scenario was proposed for the Churchill River area (Delaney, 1987). Volcanic rocks of the Transition unit may be younger than those in the PLVB. Locally interlayered mafic volcanic and pelitic rocks at the eastern margin of the Transition unit, coarse clastic rocks of the metasedimentary sequence confined to this area and predominantly pelitic rocks to the east may indicate passage into a deeper water sedimentary basin. This may suggest the presence of local submarine fans near a volcanic edifice which separated the Ourom Group on the west from Sedimentary rocks on the east.

Plastic strain features indicating oblique northward transport of an eastern block are best developed in the Transition sequence along the eastern margin of the Ourom Group. Brittle features become progressively more pronounced to the west concomitant with a progressive shallowing of the lineation and decrease in metamorphic grade. Sinistral strike-slip shear is recorded adjacent to the Tabbernor Fault.

Isograds strike NNE and indicate increasing metamorphic grade to the east. Polyphase low pressure metamorphism is suggested by the occurrence of andalusite and cordierite with overgrowth textures on staurolite and garnet. The disposition of the isograds may be closely related to uplift in the Tabbernor Stain Zone, but detailed petrographic and geochronological work must be undertaken to confirm or refute this.

10. Acknowledgments

This project was financially supported by the federal component of a Canada-Saskatchewan MDA agreement. Equipment and logistics were provided by Saskatchewan Energy and Mines. Able field assistance and good company was provided by A. MacDonald. Geological interpretations benefitted from discussions with S. Hanmer and an earlier version of the manuscript was reviewed by S. Carr.

11. References

Berthe, D., Choukroune, P. and Jegouzo, P. (1979): Orthogneiss, mylonite and non-coaxial deformation of granites, the example of the South Armoricain shear zone; J. Struc. Geol., v1, p31-42.

Budding A.J. and Kirkland, S.J.T. (1956): The geology of the Reindeer River area; Sask. Dep. Miner. Resour., Rep. 22, 42p.

Delaney, G.D. (1987): Bedrock geological mapping, Carruthers-Uskik Lakes area (part of NTS 63M-11 and -12); *in* Summary of Investigations 1987, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 87-4, p8-17.

Delaney, G.D., Carr, S.D. and Parrish, R.R., (1988): Two U-Pb zircon ages from eastern Glennie Lake Domain, Trans-Hudson Orogen, Saskatchewan; *in* Radiogenic Age and Isotopic Studies: Report 2, Geol. Surv. Can., Pap. 88-2, p51-58.

Hanmer, S. (1986): Asymmetrical pull-aparts and foliation fish as kinematic indicators; J. Struc. Geol., v8, p111-122.

Kirkland, S.J.T. (1957): The geology of the Manawan Lake area (north half), Saskatchewan; Sask. Dep. Miner. Resour., Rep. 27, 22p.

__________ (1976): The geology of the Manawan Lake area (south half), Saskatchewan; Sask. Dep. Miner. Resour., Rep. 175, 24p.

Macdonald, R. and Broughton, P. (1980): Geological map of Saskatchewan; provisional edition, Saskatchewan Geological Survey, scale 1:1 000 000.

McCaig, A.M. (1987): Deformation and fluid-rock interaction in metasomatic dilatant shear bands; Tectonophysics, v135, p121-132.

Platt, J.P. and Vissers, R.L. (1980): Extensional structures in anisotropic rocks; J. Struc. Geol., v2, p397-410.

Ramsay J.G., Casey, M. and Kligfield, R. (1983): Role of shear in development of the Helvetic fold-thrust belt of Switzerland; Geol., v11, p439-442.

Reinhardt J. (1987): Cordierite-anthophyllite rocks from northwest Queensland, Australia: metamorphosed magnesian pelites; J. Meta. Geol., v5, p451-472.

Sibbald, T.I.I. (1978): Geology of the Sandy Narrows (east) area; Sask. Dep. Miner. Resour., Rep. 170, 49p.

Wilcox, K.H. (1990): Bedrock geology and gold metallogenesis, Nielsen Lake area, Tabbernor Fault Zone; *in* Summary of Investigations 1990, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 90-4, p74-83.

A Reconnaissance Petrographic and Isotopic Study of Fluid History in the Tabbernor Fault Zone, Saskatchewan[1]

M. P. Field[2] and R. Kerrich[2]

Field, M.P. and Kerrich, R. (1991): A reconnaissance petrographic and isotopic study of fluid history in the Tabbernor Fault Zone, Saskatchewan; *in* Summary of Investigation 1991, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 91-4.

The Tabbernor Fault Zone is a prominent continuous north-striking structure that transects the Reindeer Zone of the Trans-Hudson Orogen for approximately 600 km (Byers, 1962). Interpreted as a major crustal structure (Green *et al.*, 1985) with a dominantly left lateral displacement (Furnival, 1941; Macdonald, 1976), the Tabbernor Fault Zone separates the Glennie Domain on the west from the Kisseynew Domain and the Hanson Lake Block to the east (Lewry and Sibbald, 1977; MacQuarrie, 1979). The total displacement vector, however, remains unknown. It has been recognized that deep crustal structures may act as conduits for ascending metal laden fluids that mineralize second order splays, and may also be sites for downward infiltration (Kerrich, 1989). Characterizing the fluids which migrate through major structures such as the Tabbernor Fault Zone may aid in interpretation of the fluid advection history relative to the deformational evolution and potential for mineralization.

1. Geology of the Area

The area considered for this study was recently mapped by Wilcox (1990) who divided the terrane into four distinct rock units: 1) a tonalitic intrusion, 2) intermediate to felsic volcanic to volcaniclastic rocks and arenite, 3) petite to psammopelite, and 4) highly strained rocks of unknown protolith (Figure 1). This study concentrated on Wilcox's highly strained rocks and the eastern part of the tonalite intrusion as they are adjacent to the Tabbenor Fault Zone (TBZ), were strained during development of the faults and are well exposed. The intensely strained zone typically consists of mylonite which is ultrafine grained (< 0.1 mm to 0.5 mm), with more competent relict feldspars occurring as porphyroclasts in a fine-grained chloritic schist groundmass.

A number of small discontinuous quartz veins and stringers ranging from 1 cm wide by 5 cm long to 15 cm wide and 100 cm long, cutting the mylonite and the deformed tonalite can be divided into two sets that developed sequentially:

1) early to mid-stage veins which are boudinaged, internally strained and parallel to the fabric, and
2) mid to late-stage veins which cross cut the foliation and one another comprise a second multi-stage evolution.

In general the area has a north-south-striking foliation which dips steeply to near vertical. This fabric is most dominant in the mylonite, and is evident at even greater distances from the TBZ where the pluton still possesses a weak to moderate foliation parallel to the TBZ.

2. Sample Collection and Analytical Techniques

Sample collection at Nielsen Lake concentrated on the mylonitized chloritic rock of unknown protolith (approximate UTM coordinate 619400 mE, 6104200 mN), and the eastern part of the moderately strained tonalite (approximate UTM coordinate 619550 mE, 6104800 mN). Sampling was undertaken in order to constrain the various fluid events and to evaluate the relationship of these events to adjacent wall rock. This involved sampling 1) quartz veins with different field relationships and distinct rheologies, which probably represented independent fluid events and 2) a suite from the mylonite to characterize the protolith. In addition, a number of variably-strained tonalite samples were collected, along with a small group of smokey black quartz sweats (1 cm by 10 cm) which parallel the dominant fabric in the area, to determine what effect development of the Tabbernor fault system had on the protolith.

Agate-milled whole rock powders were initially analyzed for major and selected trace elements by standard X-ray flourescence spectrometry and a suite of 33 trace elements were analyzed using ICP-MS. Oxygen isotope compositions were determined using the standard BrF_5 technique of Clayton and Mayeda (1963). Hydrogen isotope compositions were measured using a technique described by Bigeleisen *et al.*, (1952), as modified by Kyser and O'Neil (1984). All stable isotope analyses were conducted using conventional isotope ratio mass spectrometry on a Finnigan MAT 251 mass spectrometer at the University of Saskatchewan and are reported in δ notation in units of per mil relative to the V-SMOW standard.

---

(1) Saskatchewan Project A.149, from which this report derives, was funded in 1991 under the Canada-Saskatchewan Partnership Agreement on Mineral Development 1991-95 and NSERC operating grant to R.K. and NSERC infrastructure grant to R.K. and T.K. Kyser

(2) Department of Geological Sciences, University of Saskatchewan, Saskatoon, Saskatchewan

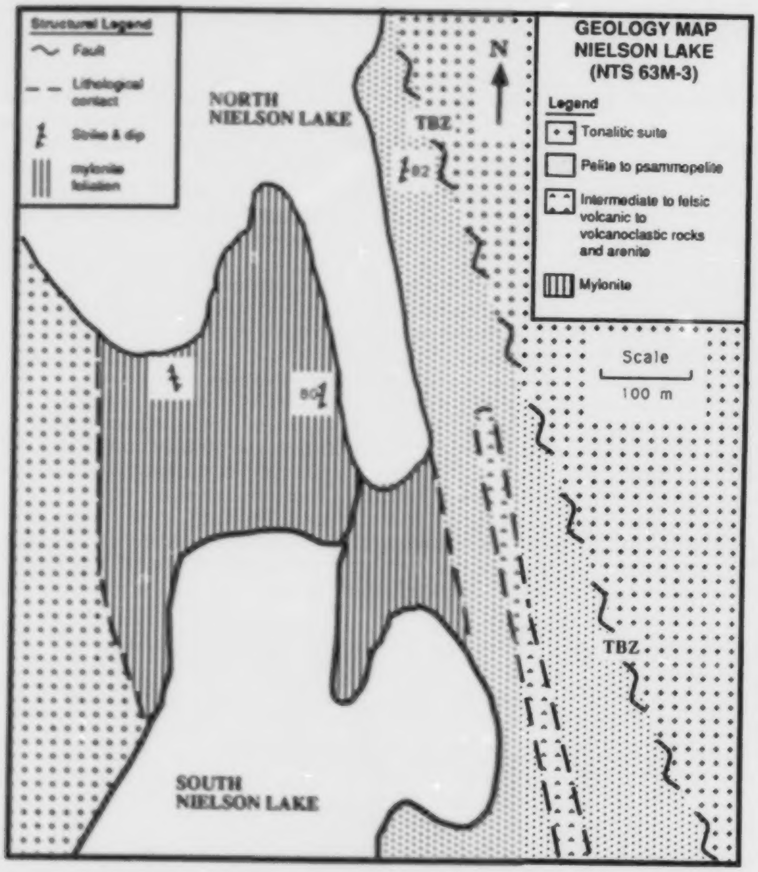


Figure 1- Geological map of the Nielson Lake area after Wilcox (1990). Tabbemor Fault Zone = TBZ.

3. Host Rock Types

a) Hand Specimen Description and Petrography

Provisional sample descriptions are based on field observations, followed by the selection of a representative suite for more detailed petrography, major element, trace element and isotopic analyses.

Tonalite

The fabric in the strained medium-grained, biotite tonalite is defined by preferred dimensional alignment of biotite. Plagioclase feldspar, biotite, quartz, sphene, and apatite are the most abundant minerals, but minor fine-grained potassium feldspar also occurs (~ 0.25 mm). Large rotated plagioclase porphyroblasts are abundant and typically have pressure shadows infilled with fine-grained quartz exhibiting undulose extinction. Pleochroic straw yellow to dark brown biotite wraps around the larger plagioclase. Sphene and apatite are generally associated with biotite.

Mylonite

In hand specimen the mylonite is a dark green, intensely deformed rock with relict feldspar porphyroclasts in a fine-grained matrix (< 0.1 mm). Petrographically, chlorite and/or actinolite-epidote-quartz-feldspar-apatite, minor sphene and ilmenite are the most abundant minerals (> 1 percent). Grains are generally less than 0.5 mm in size. Chlorite is moderately pleochroic whereas actinolite is weakly pleochroic. Quartz subgrains characterized by undulose extinction are more reduced in grain-size than the larger rotated plagioclase crystals and feature microstructures indicative of dynamic recrystalization. Compositional segregation banding of felsic-rich (quartz and feldspar) layers separated by fibrous aligned chlorite or aligned elongate actinolite grains characteristically defines the macroscopic fabric. Epidote is intergrown and associated with chlorite, and to a lesser extent with actinolite. Sphene and ilmenite occur as inclusions within chlorite and/or actinolite. With the exception of sample NL90-13, the dominant mafic mineral is either chlorite or actinolite. In sample NL90-13 chlorite occurs after actinolite, suggesting that chlorite is a retrograde product of actinolite and local synkinematic metamorphic conditions reached amphibolite grade in the mylonite.

b) Major and Trace Element Whole Rock Geochemistry

Tonalite

The least deformed part of the tonalite pluton is relatively compositionally homogeneous, with the exception of aluminum and potassium. Major oxides TiO_2, Al_2O_3, Fe_2O_3, MnO, MgO, CaO, K_2O and P_2O_5 which diminish in amount with respect to increasing SiO_2 (58.4 to 65.9 wt. percent), high K_2O values (3.5 to 4.5 wt. percent), reflect abundant biotite. The scatter in P_2O_5 corresponds to variable modal abundance of apatite.

Compositionally, this rock is too rich in K_2O (3.5 to 4.5 wt. percent) and poor in CaO (3.1 to 3.9 wt. percent) to be properly termed a tonalite, which typically have K_2O < 2 wt. percent, and CaO averages 5.4 wt. percent; rather it should be called a granodiorite or monzodiorite (Hyndman, 1985). The potassium content may not be primary, and K_2O may have been metasomatically introduced. The uniform K_2O contents argue against this.

The granodiorite features light REE enrichment, a negative Eu anomaly, and heavy REE depletion.

Mylonite

The mylonite is compositionally heterogeneous. For example, compositional ranges are: SiO_2 55.0 to 74.0 wt. percent, Al_2O_3 11.0 to 15.5 wt. percent, CaO 1.0 to 8.0 wt. percent, and Fe_2O_3 5.0 to 14.0 wt. percent. The mylonite can be subdivided into two distinct relatively homogeneous groups, A and B. Group A (NL-90-01, 02, 03, and 08) has systematically lower content of TiO_2, Fe_2O_3, MgO, Sc, and higher Zr and Hf, relative to group B (NL-90-12, 13, 14, 26). Plots of Ti vs. Zr, Ti/Zr vs. Al_2O_3/TiO_2 and Zr/Ti vs. Nb/Y (Figure 2) further emphasize this grouping. These plots also show the distinctions of the tonalite, which could not have been a precursor to the mylonite. The discrimination plots have been cast in co-ordinates of relatively immobile elements. Chondrite-normalized rare earth element (REE) patterns of the mylonite are relatively flat (Figure 3). Group A has a distinctive negative Eu anomaly and coherent patterns, whereas group B does not feature Eu anomalies and is characterized by variable light REE abundances and therefore a range of La/Yb ratios. The mylonite REE distributions are also distinct from the tonalite REE patterns.

In order to identify possible protoliths of the mylonite, the REE were normalized to an Archean trondhjemite, a Phanerzoic quartz-poor greywacke, and extended element plots were normalized to the adjacent tonalite. From Figure 4 it is clear that the protolith was neither a typical trondhjemite or greywacke, nor the adjacent tonalite.

Although the two mylonite groups clearly have two distinctive precursors, it is difficult to assess what these were. However, the remarkable similarities of their REE patterns to arc andesites (group A) and basalts (group B) coupled with chemical signatures of weathering (variability depleted CaO, Na_2O, Sr, and Ba contents) are consistent with first-cycle arc volcanoclastic sediments derived from andesite- and basalt-dominated sources. Hydrothermal alteration of such rocks, however, cannot be ruled out as an explanation for the geochemical characteristics. If the mylonites have a volcanogenic sedimentary protolith, then it may not be coincidental that deformation was concentrated in such relatively weak rock types.

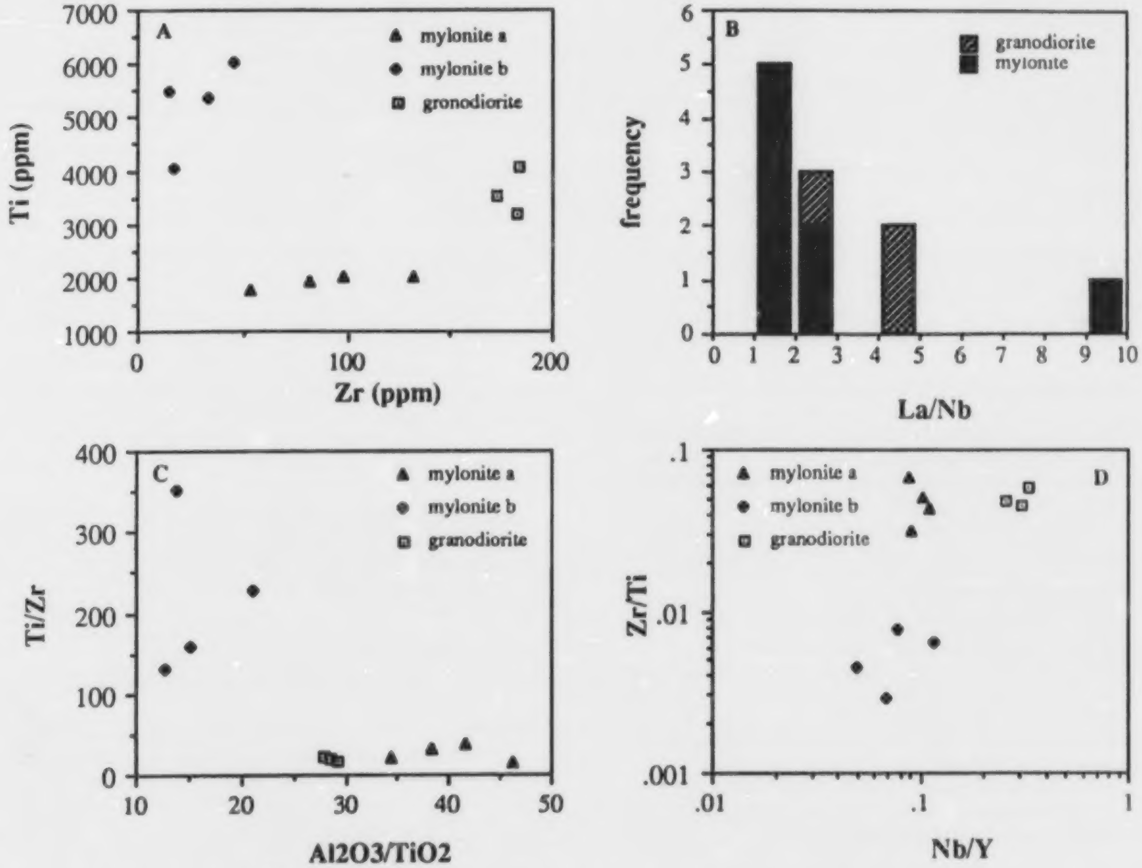


Figure 2 - Plots of relatively immobile elements discriminate between the two mylonites and the granodiorite (tonalite).

4. Quartz Veins

a) Field Descriptions, Relationships and Petrography

Evidence for fluid activity in the TBZ consists almost entirely of sets of nonmineralized quartz veins. Veins were classified in the field as discontinuous, boudinaged, early- to mid-stage conformable syndeformational, or as undeformed mid- to late-stage, post-deformational (1 cm wide continuous) cross-cutting ductile fabrics. Vein petrography reveals that the later veins are coarse grained (~5 mm diameter), with relatively undeformed anhedral crystals, whereas earlier quartz veins consist of fine-grained (~l mm), deformed, elongate grains which exhibit undulose extinction, grain size reduction and serrated grain boundaries.

b) Stable Isotope Analyses

Oxygen isotope compositions obtained from various quartz veins and quartz separates from whole rocks are given in Table 1 and plotted on Figure 5. From the veins in the mylonite it is possible to discern three distinct groupings of $\delta^{18}O$: 1) at 10.4 per mil and 10.7 per mil, 2) an intermediate population between 7.6 per mil and 9.2 per mil, and 3) very low $\delta^{18}O_{qtz}$ values between 3.6 per mil and 4.1 per mil (Table 1). The classification based on $\delta^{18}O_{qtz}$ values corresponds to interpretations of different veining events based on field relationships,

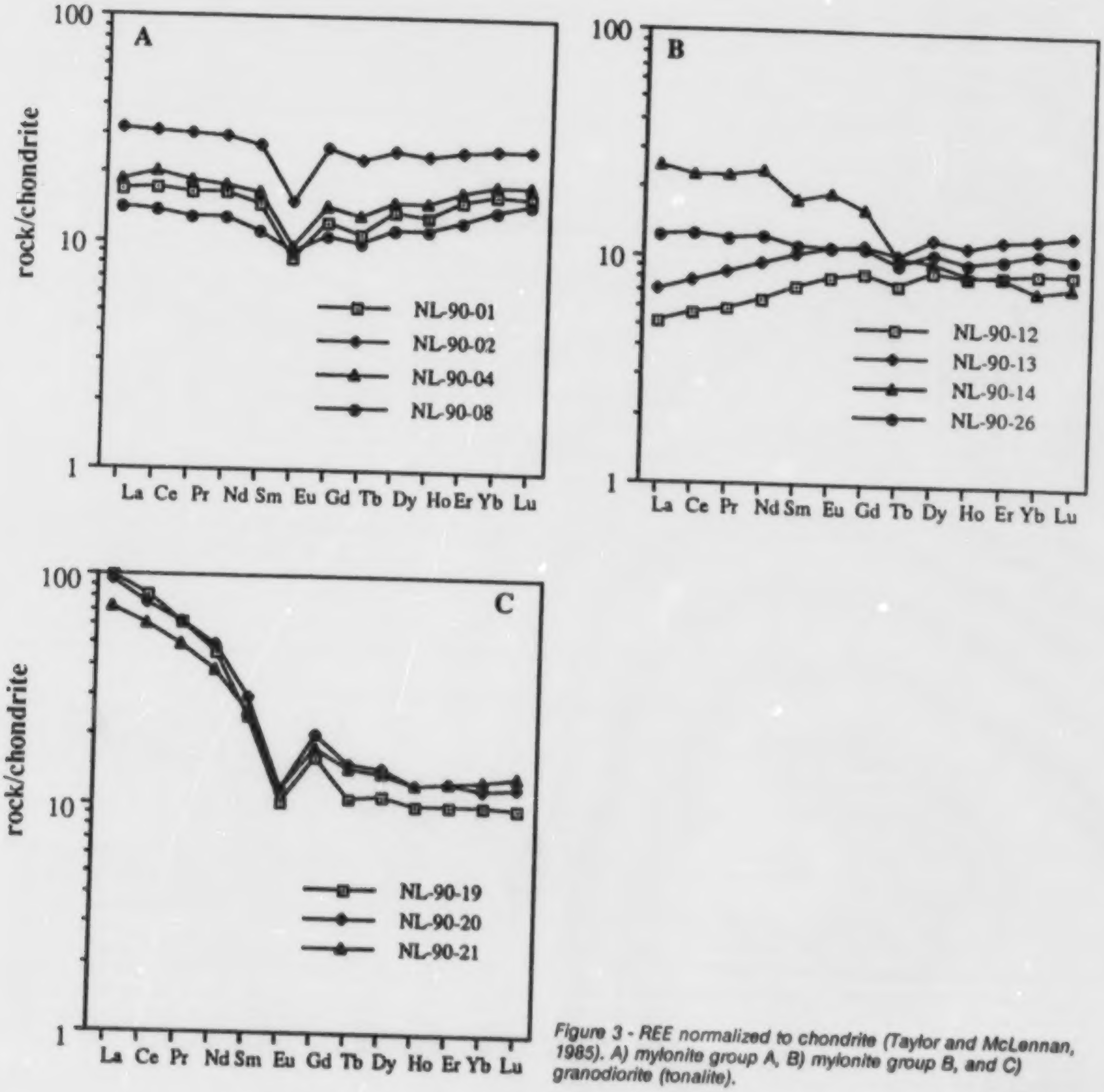


Figure 3 - REE normalized to chondrite (Taylor and McLennan, 1985). A) mylonite group A, B) mylonite group B, and C) granodiorite (tonalite).

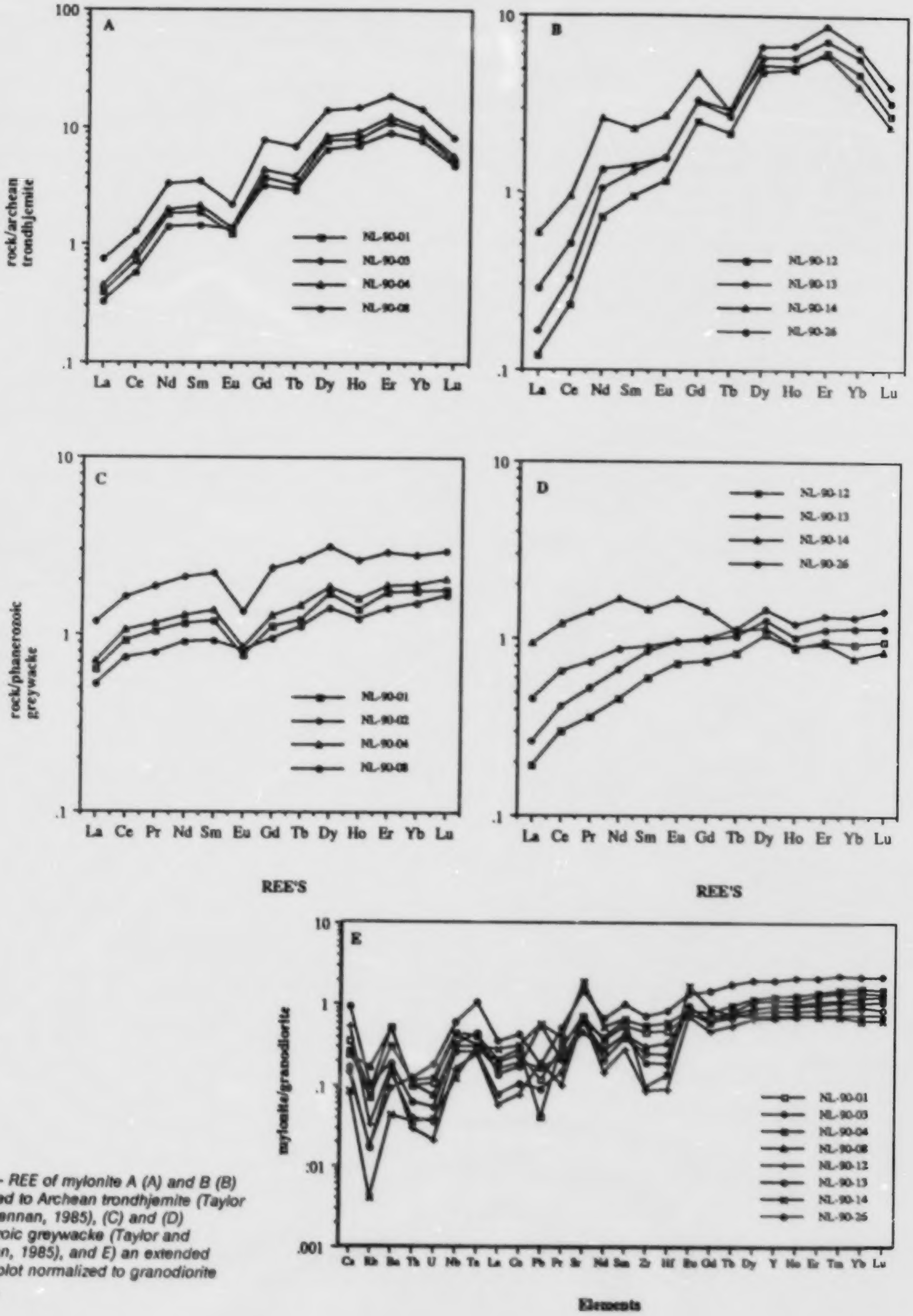


Figure 4 - REE of mylonite A (A) and B (B) normalized to Archean trondhjemite (Taylor and McLennan, 1985), (C) and (D) Phanerozoic greywacke (Taylor and McLennan, 1985), and E) an extended element plot normalized to granodiorite (tonalite).

Table 1 - Quartz $\delta^{18}O$ Values Compared to Vein Petrography and Vein Genesis (as interpreted from field relationships).

Sample #	Vein petrography	$\delta^{18}O$(per mil)	Vein genesis
Mylonite host			
NL90-70	coarse (~5mm) undeformed euhedral crystals	3.6	mid to late qtz vein
NL90-27	coarse (~5mm) undeformed euhedral crystals	3.9	mid to late qtz vein
NL90-06	coarse (~5mm) undeformed euhedral crystals	4.1	mid to late qtz vein
NL90-09	coarse (~5mm) undeformed euhedral crystals	3.9	mid to late qtz vein
NL90-10	coarse (~5mm) undeformed euhedral crystals	4.0	mid to late qtz-epi vein
* NL90-10	coarse (~5mm) undeformed euhedral crystals	-2.5	mid to late qtz-epi vein
NL90-11	fine (~1mm) deformed elongate crystals	7.6	early to mid qtz vein
NL90-08	fine (~1mm) deformed elongate crystals	7.7	early to mid qtz vein
NL90-03	fine (~1mm) deformed elongate crystals	9.2	early to mid qtz vein
NL90-13	fine (~1mm) deformed elongate crystals	8.9	early to mid qtz vein
NL90-24	fine (~1mm) deformed elongate crystals	10.4	earliest qtz vein
NL90-25	fine (~1mm) deformed elongate crystals	10.7	earliest qtz vein
NL90-02	fine (~0.5mm) deformed crystals	6.5	whole rock qtz
NL90-02	fine (~1mm) deformed elongate crystals	6.8	adjacent qtz vein
Tonalite host			
NL90-21	fine (~0.5mm) deformed crystals	10.2	whole rock qtz
NL90-21	fine (~1mm) deformed elongate crystals	8.4	smokey qtz lens
NL90-20	fine (~1mm) deformed elongate crystals	9.5	smokey qtz lens
NL90-18	coarse (~5mm) undeformed euhedral crystals	5.1	late qtz-feld vein

* vein epidote value

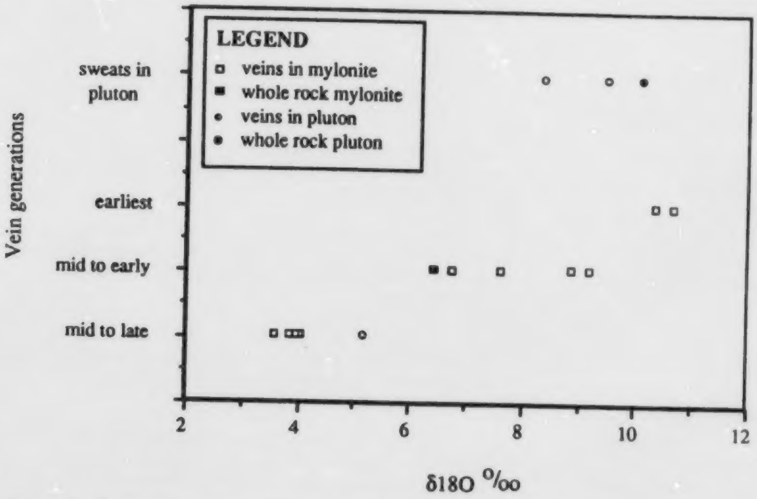


Figure 5 - Graphical representation of vein generations and their respective $\delta^{18}O$ values.

timing, and petrography (Table 1). These data suggest that at least two and possibly three, distinct fluids were involved in the evolution of the TBZ at different tectonic stages.

Epidote from a quartz-epidote vein in sample NL90-10 yielded a $\delta^{18}O$ value of -2.5 per mil (Table 1). A fluid temperature of 215°C was calculated using the quartz-epidote mineral pair (Matthews *et al.*, 1983). The $\delta^{18}O_{H_2O}$ values independently determined using epidote and quartz yield values near -6 per mil.

Earlier veins do not contain mineral pairs which may be used to determine temperature, or therefore $\delta^{18}O_{H_2O}$. Accordingly, based on petrography, temperatures appropriate to upper greenschist to lower amphibolite grade metamorphism (<550°C) were assumed for earlier veins. Using $\delta^{18}O_{qtz}$ values (Table 1), $\delta^{18}O_{H_2O}$ ranges for estimated temperatures can be determined. The earliest fluid has a $\delta^{18}O_{H_2O}$ value of about 6.5 to 8.0 per mil, consistent with a rock-buffered metamorphic or magmatic fluid. The latest fluid, corresponding to post ductile shear in the cross-cutting veins has low $\delta^{18}O_{H_2O}$ values of approximately 5.3 per mil, which may indicate that late meteoric waters infiltrated the fault zone. The intermediate group (-1.0 to 1.0 per mil and 3.0 to 4.5 per mil) may have derived by: 1) mixing of a metamorphic/magmatic reservoir with meteoric waters, 2) partial re-equilibration of meteoric fluids with the wall rock under low water rock ratios, or 3) some combination of the above.

The small syn-deformational quartz sweats have $\delta^{18}O_{qtz}$ values close to that of the pluton (Table 1), suggesting that they were probably derived from an external fluid that had equilibrated with the pluton under low water/rock ratios. In contrast, the intermediate and late stage veins clearly formed under open system conditions and higher water/rock ratios, signifying that the Tabbernor Fault Zone was in hydrologic connection with a large external fluid reservoir.

5. Conclusions

1) Petrographic observations indicate that peak pressures and temperatures during ductile deformation in mylonite of the Tabbernor Fault Zone reached amphibolite grade (<550°C), although most of the actinolite in the zone has since retrograded to chlorite as a result of lower temperatures.

2) Three distinct quartz vein generations may be defined based on field relationships, petrography, and $\delta^{18}O_{qtz}$; i) earliest magmatic/metamorphic, ii) early to mid veins with some meteoric component, and iii) late meteoric.

3) The "tonalite" is compositionally homogeneous and is more properly termed a granodiorite or monzodiorite.

4) The mylonite can be divided into two compositionally homogeneous groups. REE distributions and immobile element patterns are consistent with volcanoclastic sediments derived from an arc source.

5) On geochemical grounds it is unlikely that the tonalite was the dominant protolith to the mylonite.

6) A late quartz-epidote vein yields fluid temperature of 215°C and $\delta^{18}O$ H_2O of -6 per mil.

7) Fluid activity in the Tabbernor Fault Zone evolved from high temperature rock-buffered conditions at low water/rock ratios, to low temperature open system conditions involving infiltration of meteoric water, under brittle conditions and likely lower pressure.

6. Acknowledgments

This project was funded by SEM, an NSERC operating grant to R.K., and an NSERC infrastructure grant to R. Kerrich and T.K. Kyser. We thank K. Wilcox and R. Macdonald for geological guidance and discussions.

7. Selected Bibliography

Bigeleisen, J., Pearlman, M.L. and Prosser, H.C. (1952): Conversion of hydrogenic materials for isotopic analysis; Anal. Chem., v25, p1356 -1357.

Byers, A.R. (l962): Major faults in western part of Canadian Shield with special reference to Saskatchewan; *in* Stevenson, J.S. (ed.), Tectonics of the Canadian Shield, R. Soc. Can., Spec. Publ. 4, p40-59.

Clayton, R.N. and Mayeda, T.K. (1963): The use of bromine penta flouride in the extraction of oxygen from oxides and silicates for isotopic analysis; Geochim. Cosmochim. Acta, v27, p43-52.

Clayton, R.N., O'Neil, J.R. and Mayeda, T.K. (1972): Oxygen isotope exchange between quartz and water; J. Geophys. Res., v77, p3057-3067.

Furnival, G.M. (1941): Porcupine River, northern Saskatchewan; Geol. Surv. Can., Map 658A with marginal notes.

Graham, C.M., Sheppard, S.M.F. and Heaton, T.H.E. (1980): Experimental hydrogen stable isotope studies: I. Systematics of hydrogen isotope fractionation in the systems epidote-H_2O, zoisite-H_2O, and AlO(OH)-H_2O; Geochim. Cosmochim. Acta, v44, p128-138.

Green, A.G., Hajnal, Z. and Weber, W. (1985): The evolutionary model of the western Churchill Province and western margin of the Superior Province in Canada and the north central United States; Tectonophysics, v116, p281-322.

Hyndman, D.W. (1985): Petrology of Igneous and Metamorphic Rocks; McGraw-Hill Book Company, 307p.

Kerrich, R. (1989): Geochemical evidence on the sources of fluids and solutes for shear zone hosted mesothermal gold deposits; *in* Bursnall, J.T. (ed.), Mineralization and Shear Zones, p129-l98.

Kerrich, R. and Hyndman, D. (1986): Thermal and fluid regimes in Bitterroot Lobe-Sapphire Block detachment zone, Montana: Evidence from $^{18}O/^{16}O$ and geologic relations; Geol. Soc. Am. Bull., v97, p147-155.

Kyser, T.K. (l987): Equilibrium fractionation factors for stable isotopes; *in* Kyser, T.K. (ed.), Stable Isotope Geochemistry of Low Temperature Fluids, Miner. Assoc. Can., Short Course, v13, p1-84.

Kyser, T.K. and O'Neil, J.R. (l984): Hydrogen isotope systematics of submarine basalts; Geochim. Cosmochim. Acta, v48, p2123-2133.

Lewry, J.F. and Sibbald, T.I.I. (1977): Variation in lithology and tectonometamorphic relationships in the Precambrian basement of northern Saskatchewan; Can. J. Earth Sci., v14, p1453-1467.

Macdonald, R. (l976): Compilation geology: Pelican Narrows (63M) and Amisk Lake (63L); *in* Summary of Investigations 1976; Saskatchewan Geological Survey, Sask. Dep. Miner. Resour., p53-57.

__________ (1987): Update on Precambrian geology and domainal classifications of northern Saskatchewan; *in* Summary of Investigations 1987, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 87-4, p87-104.

MacQuarrie, R.R. (1979): Geological re-investigation mapping, Birch Portage south (NTS area 63L-15(s)); *in* Summary of Investigations 1979, Saskatchewan Geological Survey, Sask. Dep. Miner. Resour., Misc. Rep. 79-10, p29-38.

Matthews, A., Goldsmith, J.R. and Clayton, R.N. (1983): Oxygen isotope fractionations between zoisite and water; Geochim. Cosmochim. Acta, v47, p645-654.

Padgham, W.A. (1968): The geology of Deschambault Lake district, Saskatchewan; Sask. Dep. Miner. Resour., Rep. 114, 92p.

Taylor, S.R. and McLennan, S.M. (1985): The Continental Crust: its Composition and Evolution; Blackwell Scientific Publications.

Wilcox K. (1990): Bedrock geological mapping, Nielson Lake area, Tabbernor Fault Zone; *in* Summary of Investigations 1990, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 90-4, p74-83.

C. Contributed Papers, Northern Studies

The Lithoprobe Trans-Hudson Orogen Transect (THOT): Status Report

John F. Lewry[1] *and Z. Hajnal*[2]

Lewry, J.F. and Hajnal, Z. (1991): The Lithoprobe Trans-Hudson Orogen Transect (THOT): status report; *in* Summary of Investigations 1991, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 91-4.

Full operational funding of the Lithoprobe-sponsored Trans-Hudson Orogen Transect (THOT) in Saskatchewan and Manitoba began in 1991. Broad objectives of the project are to increase understanding of the geodynamic evolution of this Early Proterozoic collisional orogen, and the processes involved therein, by determining the 3-D crustal expression and other characteristics of its lithotectonic elements. More specific major objectives include:

1. to constrain the subsurface location of bounding Archean cratonic margins and document the extent and thickness of Archean crust beneath allochthonous terranes of the internal Reindeer Zone (cf. Lewry *et al.*, 1990);
2. to determine the degree to which apparent northwesterly subduction polarity which prevailed during orogenic convergence, and resultant gross orogenic asymmetry seen at current erosion levels, are reflected in deep crustal structure;
3. to determine 3-D geometry of major faults and shear zones within and bounding lithotectonic domains, including those at the Churchill-Superior boundary, and the Snowbird line on the northwest side of the orogen;
4. to test the model of major thrust/nappe crustal imbrication in the Reindeer Zone (Lewry *et al.*, 1990; Zwanzig, 1990);
5. to determine spatial and 3-D structural relationships of the North American Central Plains conductivity anomaly (NACP) to major lithotectonic orogenic elements; and
6. to compare and contrast crustal structure of the Trans-Hudson Orogen with Phanerozoic collisional zones, such as the Himalayas and Appalachians.

The scientific programme is centred about seismic reflection (vibroseis) profiling along one main, and two subsidiary transect corridors which allow investigation of crustal structure across the entire orogen. The Main Transect corridor runs approximately 1200 km from central Manitoba to west-central Saskatchewan (following Highways 373, 6 and 10, in Manitoba, and 106, 165, 965 and 153, in Saskatchewan) and crosses most major lithotectonic elements of the orogen. Secondary corridors extend 240 km along the Wollaston Lake road (Highway 105), in Saskatchewan, and 410 km along the Thompson-Lynn Lake-Reindeer Lake line (Highways 394, 391 and connecting access roads).

The reflection seismic programme will be supported by a full range of related multidisciplinary geoscience investigations, including seismic refraction, gravity, magnetic and electromagnetic surveys, investigation of physical rock properties and potential field modelling, paleomagnetic, petrological, structural, geochemical, isotopic and geochronological studies along and adjacent to the transect corridors. These will involve at least forty university geoscientists. Coincidentally, revision bedrock geological mapping and other studies planned by the Saskatchewan Geological Survey, Manitoba Geological Service branch and the Geological Survey of Canada are being concentrated in the vicinity of transect corridors. Total anticipated funding for the overall program, from all sources over the life of the THOT project, is ca. $7 to 8 million.

In the summer of 1991 the first phase of reflection seismic data collection along the Main Transect corridor was completed and a number of other Lithoprobe-funded research projects were begun or continued from last year. These are summarized in the following sections.

1. Reflection Seismic (Vibroseis) Profiling

Between July 20 and September 18, 1991, twelve reflection seismic (vibroseis) profile lines totalling 1,070 km were surveyed along the eastern part of the Main Transect corridor in Saskatchewan and Manitoba. This represents the largest continuous data acquisition programme undertaken by Lithoprobe in a single operational season. The survey used a 240 channel, MDS-18 telemetric recording system and four Hemi-50 vibroseis units. The maximum 80 ton peak load force generated by the latter represents the largest input energy source so far deployed by a Lithoprobe survey. The 24 hours-per-day operation and high efficiency of the computerized telemetry system resulted in a record 20 km average daily survey coverage. Excellent field-record quality was maintained throughout. This first phase of

---

(1) Department of Geology, University of Regina, Regina, Saskatchewan.
(2) Department of Geological Sciences, University of Saskatchewan, Saskatoon, Saskatchewan; Co-transect leaders.

the program was designed to examine several key orogenic segments in the eastern part of the orogen.

Line 1a (a 20 km high-resolution line) and **Line 1b** (a contiguous 40 km regional survey line) were run mainly along Highway 391, starting in the Thompson, Manitoba, city limits. The lines cross the contact between the Eastern (Churchill-Superior) Foreland Zone and the internal Reindeer Zone. International Nickel Company provided funding for the high resolution line.

Line 2, runs along Highway 373 from the Highway 6 intersection, terminating east of Jenpeg, Manitoba. It crosses part of the Superior craton and the Thompson belt, to the contact between the latter and the Reindeer Zone.

Line 3 begins at Clarke Lake and runs along Manitoba Highways 6 and 39 to the intersection of Highways 39 and 10. Again, it crosses part of the Foreland Zone and extends across the northern part of the Flin Flon-Snow Lake Domain.

The short north-south **Line 4** runs along Highway 392 to the Snow Lake town limits. The orientation of this profile, and its connection with line 3, was planned to explore the 3-D structural geometry of the north-central part of the Flin Flon-Snow Lake Domain.

Line 5 and the connecting **Line 6** follow Manitoba Highway 10, through the city of Flin Flon to the Saskatchewan-Manitoba border, traversing a structurally complex part of the Flin Flon Domain.

Line 7 comprises three segments: one from Highway 10 to the southern tip of Barret Bay, another starting at the south end of Syme Lake and running through the town of Sherridon, Manitoba to end northeast of Narrows Lake, the third extending approximately 20 km north-northeast from the east end of Batty Lake along a forestry road. This approximately north-south, essentially continuous profile crosses the Flin Flon-Kisseynew domainal boundary zone.

A planned low-priority **Line 8** was omitted due to funding constraints.

Line 9 begins at Creighton, Saskatchewan, and continues west along Highways 106 and 165 to the Highway 2 intersection. This long survey line extends across the Flin Flon and Glennie Domains, crossing both the Tabbernor Fault Zone and the projected northerly extension of the NACP conductivity anomaly.

Line 10 runs along Saskatchewan Highway 135 from its intersection with Highway 106 north and northeastward to Sandy Bay, crossing the Hanson Lake Block, including the Sahli Archean inlier, Early Proterozoic allochthons overlying the 'Pelican Slide' (Shi and Lewry, this volume) and the Hanson Lake-Kisseynew boundary to the north.

Line 11 follows Highway 2 to its southern intersection with Highway 165, crossing the contact between the Glennie and La Ronge-Rottenstone Domains.

Line 12 follows the western part of Highway 165 to its intersection with Highway 914. This profile covers part of the Rottenstone Domain, the Wathaman Batholith and Needle Falls Shear Zone, ending in the eastern Wollaston Domain.

Two ancillary experimental investigations formed a significant part of the 1991 field programme:

1) Three expanded spread ('walk-away') surveys, in the Superior craton, Flin Flon Domain and Glennie Domain, were conducted in order to extract more accurate crustal velocity/depth functions, thus providing high-precision data for development of analytical procedures to obtain high-quality acoustic properties from deep-sounding reflection data.

2) A concurrent 200 km dynamite profile, covering all of lines 11 and 12 and a western portion of line 9 (i.e. from western Glennie Domain to the Wollaston Domain). This special, explosives source-based data acquisition should allow examination of deeper parts of the crust and will provide information for more detailed delineation of major lithotectonic domain-bounding shear zones in this central part of the orogen.

2. Summary of Related Geoscience THOT Projects Funded for 1991-92

The Lithoprobe NSERC Committee received 24 proposals in 1991-92 totalling $532,609.00 for research related to the Trans-Hudson Orogen Transect. Thirteen of these were funded, for a total of $224,000.00, comprising about a quarter of total Lithoprobe transect funding in the year. Two projects were funded for two years (into 1992-93) for an additional $79,000.

The following list summarizes 1991-92 THOT related-geoscience projects approved in 1991-92 for Saskatchewan and Manitoba. Initial results of these investigations and of the reflection seismic profiling will be presented at the THOT Lithoprobe workshop to be held in late March or early April 1992.

Table 1 - Lithoprobe-NSERC Sponsored Research Projects Related to the Trans-Hudson Orogen Transect.

In Saskatchewan:

Project Title	Personnel Involved
1. P-T-t metamorphic and fluid evolution, and age constraints on evolution of the Pelican Slide high-strain zone, Hanson Lake Block: implications for development of Trans-Hudson lithosphere	R. Kerrich, T.K. Kyser, M.R. Stauffer, Min Sun (U. of Saskatchewan), G. Edwards (Athabasca U.) and J.F. Lewry (U. of Regina)
2. Origin, age and alteration history of the Boundary Intrusions and mafic-ultramafic intrusions in the Flin Flon domain	T.K. Kyser and D.S. O'Hanley (U. of Saskatchewan)
3. Petrogenesis of the Wathaman Batholith (Reindeer Lake region)	T.K. Kyser and M.R. Stauffer (U. of Saskatchewan)
4. Static Physical Properties of rocks in the Trans-Hudson Orogen	D.R. Schmitt and T. Chacko (U. of Alberta)
5. Structural analysis of the MacLennan Lake Tectonic Zone	W.M. Schwerdtner and graduate student (U. of Toronto)
6. Physical properties of rocks from the Trans-Hudson Orogen	D. Stead, C.M.R. Fowler, M.J. Reeves and E.G. Nisbet (U. of Saskatchewan)
7. Measuring the closure of the Trans-Hudson Orogen by paleomagnetism	D.T.A. Symons (U. of Windsor)
In Manitoba:	
8. Thermal/Metamorphic History of the Reindeer Zone	T.M. Gordon (U. of Calgary)
9. Geochemical and isotopic characterization of anorogenic alkaline magmatism in the Reindeer Zone	N.M. Halden (U. of Manitoba) and B. Fryer (Memorial U.)
10. Paleomagnetism of the Pikwitonei granulite domain	H.C. Halls (U. of Toronto)
11. U-Pb geochronology of the southern Trans-Hudson Orogen in Manitoba	N. Machado and J. David (UQAM)
12. Deformation history and kinematic analysis of the Snow Lake/Kisseynew domain boundary	P.F. Williams and post-doctoral student (U. of New Brunswick)
General:	
13. Rheology of lower continental and oceanic crust	J.V. Ross (UBC)

3. References

Lewry, J.F., Thomas, D.J., Macdonald, R. and Chiarenzelli, J. (1990): Structural relations in accreted terranes of the Trans-Hudson Orogen, Saskatchewan: telescoping in a collisional regime?; *in* Lewry, J.F. and Stauffer, M.R. (eds.), The Early Protereozoic Trans-Hudson Orogen of North America, Geol. Assoc. Can., Spec. Pap. 37, p76-94.

Zwanzig, H.V. (1990): Kisseynew gneiss belt in Manitoba: stratigraphy, structure and tectonic evolution; *in* Lewry, J.F. and Stauffer, M.R. (eds.), The Early Protereozoic Trans-Hudson Orogen of North America; Geol. Assoc. Can., Spec. Pap. 37, p95-120.

Detrital Zircons from Missi Metasedimentary Rocks, Flin Flon Basin: Constraints on Age and Provenance[1]

Kevin M. Ansdell[2], T. Kurt Kyser[2], Mel R. Stauffer[2] and Garth Edwards[3]

Ansdell, K.M., Kyser, T.K., Stauffer, M.R. and Edwards, G. (1991): Detrital zircons from Missi metasedimentary rocks, Flin Flon Basin: Constraints on age and provenance; *in* Summary of Investigations 1991, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 91-4.

Missi metasedimentary rocks in the Flin Flon Domain form part of a discontinuous series of molasse-type sediments which have been correlated with other psammitic gneisses (Stauffer, 1990) throughout the Reindeer Zone of the Trans-Hudson Orogen in northern Saskatchewan and Manitoba. In the Flin Flon Domain they consist of fluvial sandstones and conglomerates unconformably overlying volcanic arc and sedimentary interarc material, and are thought to have been deposited synchronously with mountain building during the Trans-Hudson orogeny.

Thus far, very few attempts have been made to constrain the age of the Missi molasse sedimentation. Delaney *et al.* (1988) determined the U-Pb ages of detrital zircon grains from the Ourom Lake Meta-arkose in the Glennie Domain, as well as the Wood Lake granodiorite which intrudes the supracrustal sequence, and suggested that the age of sedimentation was 1850 ±1 Ma. The only constraint on the timing of sedimentation in the Flin Flon Domain is provided by a U-Pb zircon age of 1832 ±2 Ma obtained from a felsic rock near Snow Lake, which is interpreted to be a rhyolitic lava interbedded with Missi Group metasedimentary rocks (Gordon *et al.* 1990). However, this age is at variance with the timing of post-Missi plutons in the Flin Flon area. The Phantom Lake granite (1840 ±7 Ma) crosscuts one of the Boundary Intrusions, which themselves crosscut Missi Group rocks, whereas the Graham Trail porphyry (1841 ±18 Ma) intrudes Missi metasedimentary rocks near the Graham gold occurrence (Ansdell and Kyser, 1990, 1991).

A study of detrital zircons from Missi metasedimentary rocks in the Flin Flon Basin, near Flin Flon, was initiated to provide further limits on the age of sedimentation, and to provide information on the age of possible source terranes. The Flin Flon Basin (Figures 1 and 2) was chosen as the study area because the stratigraphy of the rocks has been described in detail by Stauffer (1990).

1. Geological Setting of the Flin Flon Basin

The Flin Flon Basin forms part of the Flin Flon greenstone belt, which is one of the lithotectonic domains of the Reindeer Zone of the Trans-Hudson Orogen in northern Saskatchewan and Manitoba. The oldest rocks in the western part of the Flin Flon Domain (Figure 1) are termed Amisk Group, and consist mainly of tholeiitic to calc-alkaline volcanic and volcaniclastic rocks (Stauffer *et al.*, 1975). Zircons from a rhyolitic tuff yield an U-Pb age of 1886 ±2 Ma (Gordon *et al.*, 1990), whereas zircons from the South Main rhyolite yield an imprecise U-Pb age of 1925 +50/-30 Ma (Syme *et al.*, 1991). However, the total time span over which volcanic activity occurred in the western part of the Flin Flon Domain is presently unknown.

The Missi rocks, which are of interest in this study, occur in three basins in the western part of the Flin Flon Domain, namely the Flin Flon, Amisk and Athapap basins (Stauffer, 1990). A well-developed paleoregolith is present in pre-Missi rocks below the unconformity indicating extensive pre-Missi subaerial weathering, although in most places Missi rocks are in fault contact with adjacent rocks (Stauffer, 1990). The Missi sedimentary rocks in the Flin Flon Basin consist of a 2.5 km thick sequence containing two fining-upward sequences of conglomerates and sandstones (Figure 2), which are thought to represent the transition from proximal fan to braided stream deposits. Stratigraphic way-up indicators are well preserved, even though these rocks have undergone greenschist metamorphism and polyphase deformation during the Hudsonian orogeny. The deformation has, however, made determination of paleocurrent directions impossible.

A wide variety of intrusive bodies ranging in composition from gabbroic to granitic, and in age from syn-volcanic to post-tectonic crosscut these supracrustal rocks. Zircons from these intrusions yield U-Pb and $^{207}Pb/^{206}Pb$ ages ranging from 1875 to 1835 Ma (Gordon *et al.*, 1990; Ansdell and Kyser, 1991).

All rocks in the western part of the Flin Flon Domain, including the Missi sedimentary rocks in the Flin Flon

---

(1) Funded by a University (NSERC) - Industry (CAMECO) research grant

(2) Department of Geological Sciences, University of Saskatchewan, Saskatoon, Saskatchewan, S7N 0W0
(3) Geology Department, Athabasca University, Athabasca, Alberta, T0G 2R0

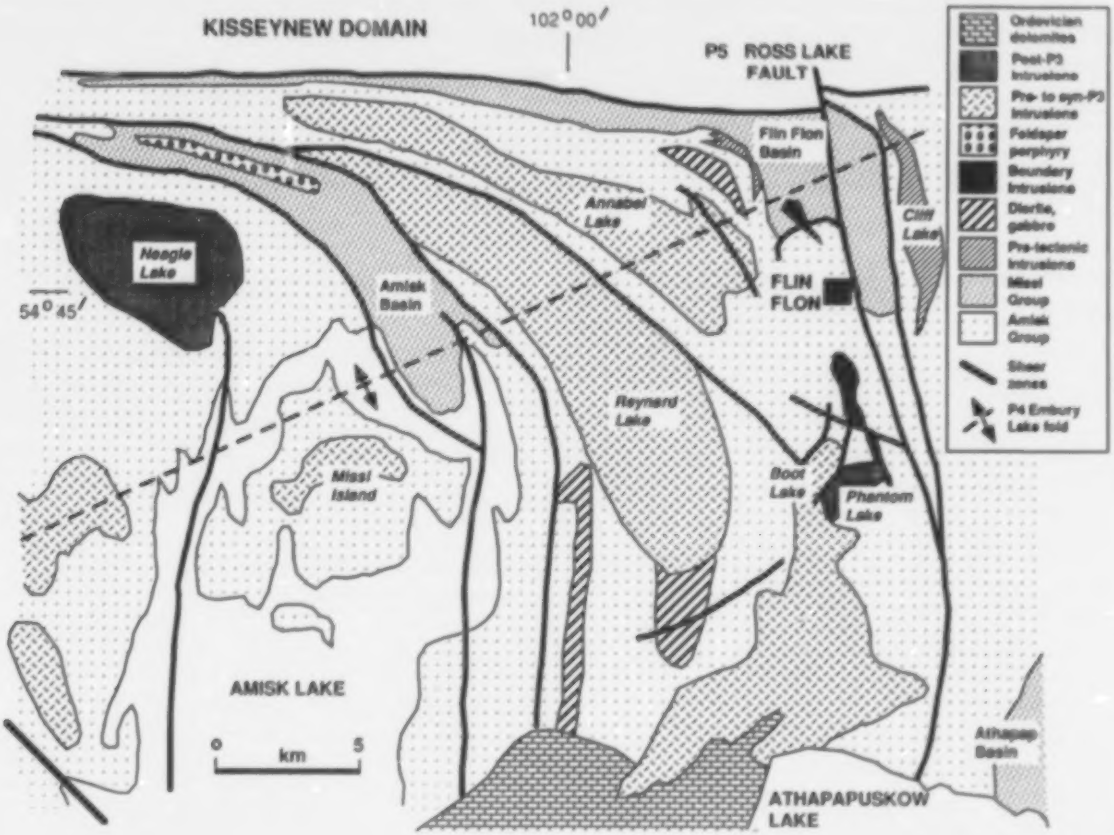

KISSEYNEW DOMAIN
102°00'
P5 ROSS LAKE FAULT
Flin Flon Basin
Annabel Lake
Cliff Lake
FLIN FLON
Neagle Lake
54°45'
Amisk Basin
Reynard Lake
Missi Island
Boot Lake
Phantom Lake
AMISK LAKE
0 km 5
Athapap Basin
ATHAPAPUSKOW LAKE
Ordovician dolomites
Post-P3 intrusions
Pre- to syn-P3 intrusions
Feldspar porphyry
Boundary intrusions
Diorite, gabbro
Pre-tectonic intrusions
Missi Group
Amisk Group
Shear zones
P4 Embury Lake fold

Figure 1 - Simplified geological map of the western Flin Flon Domain (modified after Byers and Dahlstrom, 1954; Byers et al., 1965; Stauffer, 1984), showing the location of the Flin Flon, Amisk and Athapap basins (Stauffer, 1990). Granitoids that have been dated using the single-zircon Pb-evaporation technique (Ansdell and Kyser, 1990, 1991) are also indicated.

Basin, have been variably deformed by up to five deformation events. The morphology and orientation of structures developed in the vicinity of the Flin Flon Basin are described in detail by Stauffer and Mukherjee (1971) and Fedorowich *et al.* (in press). The dominant penetrative foliation and north-south ductile shear zones developed during the third phase of deformation (P3). These shear zones were reactivated under brittle-ductile conditions during the development of the Embury Lake fold (P4). The Ross Lake fault system (P5) crosscuts the Embury Lake fold, and has a dominantly brittle character.

Regional metamorphism also post-dates deposition of Missi sedimentary rocks, and in the Flin Flon area the metamorphic grade increases from prehnite-pumpellyite grade in the southeast to amphibolite grade at the contact with the Kisseynew Gneisses (Digel and Gordon, 1991). Peak regional metamorphism is considered to be broadly synchronous with P3 and P4, and isograds are offset by P5 fault zones.

2. Sampling and Analytical Technique

Samples were taken from each of the six stratigraphic subdivisions of the Missi Group in the Flin Flon Basin (Figure 2) to determine whether there is any vertical change in age of zircons, and thus, of provenance. Zircons were separated using conventional separation techniques. Each sample consists of euhedral, subrounded and rounded zircons, and zircon fragments, and attempts were made to analyze a range in morphological type to obtain an indication of the possible range in ages of zircon. Hand-picked, non-magnetic, 125 to 250 μm zircons were analyzed using the single-zircon Pb-evaporation technique (Kober, 1987; Ansdell and Kyser, 1990, 1991).

3. Results

Euhedral and slightly rounded zircons predominate in each sample, which suggests that the majority of zircons were derived locally. Rounded zircons only represent about 10 to 20 percent of the grains in each sample but do indicate that a small proportion of the

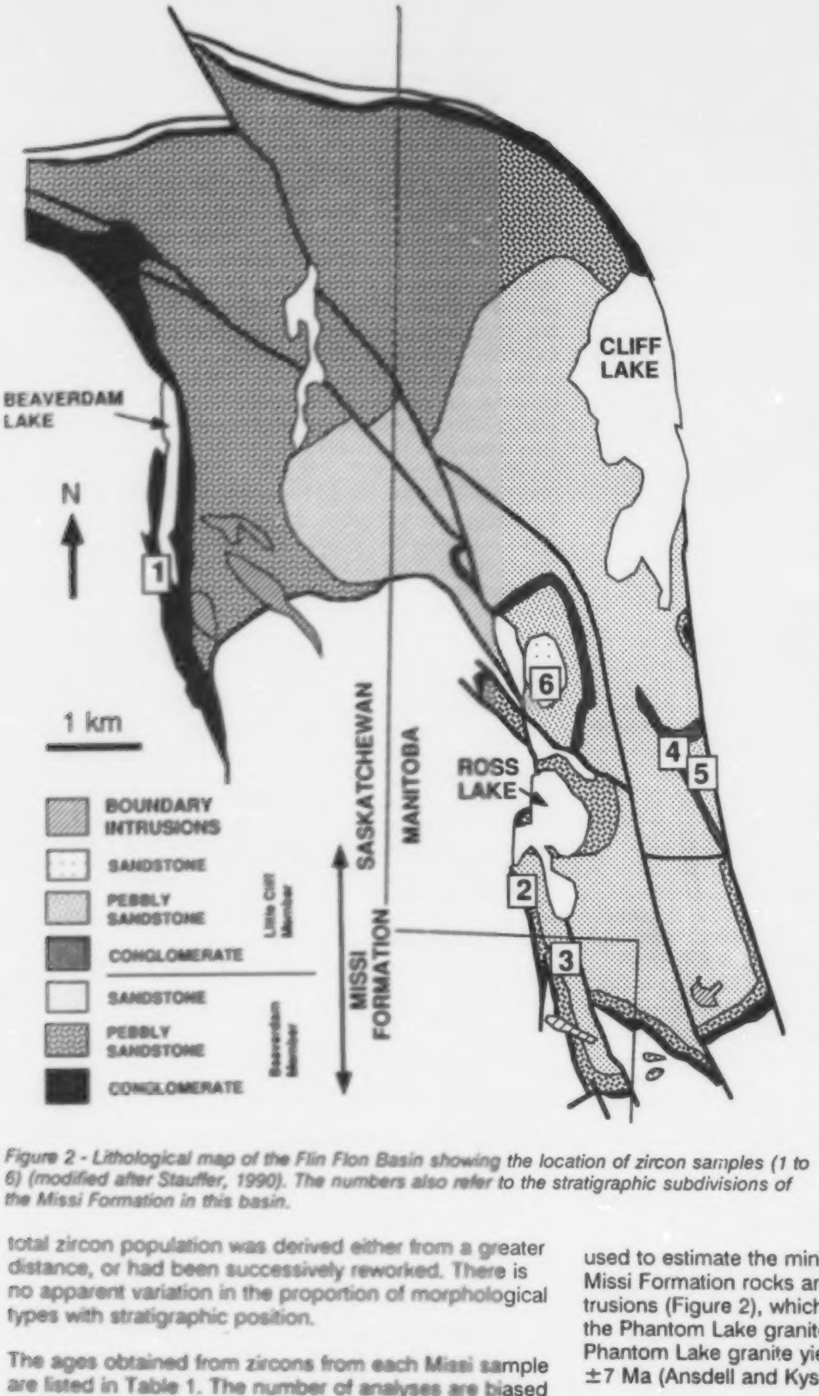


Figure 2 - Lithological map of the Flin Flon Basin showing the location of zircon samples (1 to 6) (modified after Stauffer, 1990). The numbers also refer to the stratigraphic subdivisions of the Missi Formation in this basin.

total zircon population was derived either from a greater distance, or had been successively reworked. There is no apparent variation in the proportion of morphological types with stratigraphic position.

The ages obtained from zircons from each Missi sample are listed in Table 1. The number of analyses are biased towards the less-common rounded zircons so that the full range of ages of zircons could be detected.

The youngest zircons from each stratigraphic unit range from 1854 ±13 Ma (Missi 5) to 1866 ±24 Ma (Missi 2). Based on these analyses, the youngest zircons in the lower fining-upward sequence have average ages slightly older (1860 to 1866 Ma) than the average ages of the youngest zircons in the upper fining-upward sequence (1854 to 1858 Ma), although they overlap within analytical error. Overall, the majority of zircons indicate that the dominant source terrane had an age of between about 1855 and 1900 Ma. Most of the younger zircons are euhedral or slightly rounded, although some of the rounded zircons also had a young source terrane (Table 1).

Four sub-rounded to rounded zircons yield ages older than 2000 Ma (Table 1) indicating that either a distant Early Proterozoic or Late Archean terrane was being eroded at the time of Missi deposition or that an older sedimentary source was eroded and the zircons reworked.

4. Age of Missi Sedimentation and Source of Zircons

The age of zircons within Missi Formation samples should provide an upper limit for the age of sedimentation. The youngest zircons have median ages older than 1854 (Table 1), which is about 30 Ma younger than the age of underlying Amisk volcanic rocks and indicates that Missi sediments were deposited after 1854 Ma.

The age of intrusions that crosscut the Flin Flon Basin metasedimentary rocks can be used to estimate the minimum age of sedimentation. Missi Formation rocks are intruded by Boundary Intrusions (Figure 2), which themselves are crosscut by the Phantom Lake granite (Figure 1). Zircons from the Phantom Lake granite yield a ^{207}Pb/^{206}Pb age of 1840 ±7 Ma (Ansdell and Kyser, 1990, 1991).

The data are compatible with the interpretation that Missi sediments were deposited between 1854 ±13 and 1840 ±7 Ma, and the geologic relations require that they were deposited in a much shorter time period within this range. The Missi sedimentary rocks exposed

Table 1 - Morphology and Age of Detrital Zircons from the Missi Formation in the Flin Flon Basin.

Sample	Morphology	Age (Ma ±2 sigma)
Missi 1	Euhedral	1861 ±18
	Subrounded	1869 ±53
	Euhedral	1874 ±43
	Euhedral	1914 ±6
Missi 2	Euhedral	1866 ±24
	Euhedral	1877 ±13
	Rounded	1892 ±28
	Rounded	2491 ±36
Missi 3	Euhedral	1860 ±8
	Rounded	1871 ±40
	Subrounded	1928 ±24
	Euhedral	1951 ±51
	Subrounded	2186 ±18
Missi 4	Subrounded	1856 ±19
	Subrounded	1858 ±23
	Rounded	1866 ±26
	Subrounded	1921 ±18
Missi 5	Euhedral	1854 ±13
	Euhedral	1865 ±25
	Euhedral	1881 ±8
	Rounded	2529 ±20
Missi 6	Euhedral	1858 ±8
	Euhedral	1869 ±9
	Euhedral	1879 ±9
	Subrounded	1891 ±8
	Euhedral	1896 ±26
	Rounded	2092 ±38

Location of samples indicated on Figure 2.

within the Flin Flon Basin were thus deposited before similar rocks in the Snow Lake area, which have been dated at 1832 ±2 Ma (Gordon *et al.*, 1990). This disparity in age of sedimentation emphasizes the possible diachronous and spatially irregular nature of fluvial sedimentation within this portion of the Trans-Hudson orogen, which is probably related to the likely complex surface expressions of crustal deformation events and erosion during long-lived collision within the Reindeer Zone.

The age of the zircons is assumed to be indicative of the age of the source terrane for the Missi sediments. However, the rounded zircons may indicate successive reworking, and thus they provide information on the ultimate source of the zircon and not necessarily the age of the last rock from which they were eroded.

The Missi consists of fluvial sedimentary rocks of immature character which, in conjunction with the high proportion of euhedral zircons, indicates that the predominant source is likely to be local. A number of the granitoid plutons presently exposed in the Flin Flon area have ages similar to those of the detrital zircons (e.g. Annabel Lake, 1860 ±6 Ma; Reynard Lake, 1853 ±8 Ma; Cliff Lake, 1874 +32/-25 Ma: Ansdell and Kyser, 1990, 1991; Gordon *et al.*, 1990; Figure 1), and thus represent the most likely source rocks for zircons with ages of 1854 to 1880 Ma.

Two felsic volcanic rocks from the Amisk Group in the Flin Flon area have been dated by U-Pb in zircon at 1886 ±2 Ma (Gordon *et al.*, 1990) and 1925 +50/-30 Ma (Syme *et al.*, 1991). Other U-Pb ages in the range 1880 to 1950 Ma have been obtained from tonalitic and granitic plutons and gneisses in the eastern Flin Flon Domain (1889 +8/-6 Ma; 1886 +17/-9 Ma, Bailes *et al.*, 1991), the southern Kisseynew Domain (1890 +12/-9 Ma, Gordon *et al*, 1990; 1892 +66/-25 Ma, Hunt and Zwanzig, 1990), the Glennie Domain (1893 ±35 Ma, Van Schmus *et al.*, 1987) and from the western part of the Flin Flon Domain (1908 ±3 Ma, Heaman *et al.*, this volume). These or similar rocks are possible sources for zircons in the Missi Group having ages of 1880 to 1950 Ma. The euhedral character of some zircons implies that they may have been derived from a more local source in the western Flin Flon Domain. The Mystic Lake granodiorite, recently provisionally dated by U-Pb in zircon at 1908 ±3 Ma (Heaman, this volume), may be such a source.

The source for the oldest zircons is more problematic. U-Pb zircon geochronology in the Thompson area suggests that collision between the Superior Province and the Reindeer Zone did not occur until about 1810 Ma (Machado, 1990), and so the Flin Flon area was probably separated from Superior Province Archean rocks by an ocean basin at the time of Missi sedimentation. Bickford *et al.* (1990) suggest that collision between the Reindeer Zone, and the Rae-Hearne provinces occurred earlier, at about 1855 Ma, with related intrusion of the Wathaman batholith. The older rocks to the northwest were adjacent to the present Reindeer Zone during Missi sedimentation, and represent a possible source for the well-rounded detrital zircons having the oldest $^{207}Pb/^{206}Pb$ ages. However, there are also small inliers of Archean gneisses in the Hanson Lake Block (Bell and Macdonald, 1982; Craig, 1989) and Glennie Lake Domain (Chiarenzelli *et al.*, 1987) and these, or similar rocks, exposed during Missi sedimentation are also possible sources for the zircons.

5. Conclusions

1. The age of Missi sedimentation is constrained to between the age of the Phantom Lake granite (1840 Ma), and the age of the youngest zircon (1854 Ma).

2. There is no vertical variation in the age of zircons, or zircon morphology, which indicates that there were no distinct changes in provenance during Missi deposition.

3. The lack of paleocurrent directions makes the determination of provenance difficult. The dominant range of zircon ages (1855 to 1955 Ma) is; however, consistent with derivation from early granitoids in the western Flin Flon Domain, Amisk Group felsic rocks and associated syn-volcanic granitoids. The euhedral nature of most of these zircons provide indirect evidence for the presence of pre-1900 Ma granitoids close to the Flin Flon Basin.

4. The oldest, rounded zircons are derived from an unknown source, either older Proterozoic sedimentary rocks, or a distant Archean or early Proterozoic granitoid terrane.

6. References

Ansdell, K.M. and Kyser, T.K. (1990): Age of granitoids from the western Flin Flon Domain: An application of the single-zircon Pb-evaporation technique; *in* Summary of Investigations 1990, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 90-4, p136-142.

__________ (1991): Plutonism, deformation and metamorphism in the Proterozoic Flin Flon greenstone belt, Canada: Limits on timing provided by the single-zircon Pb-evaporation technique; Geol., v19, p518-521.

Ansdell, K.M., Stauffer, M.R., Kyser, T.K. and Edwards, G. (1991): Age and source of detrital zircons from the Missi Group: A Proterozoic molasse deposit, Trans-Hudson Orogen; Geol. Assoc. Can./ Mineral. Assoc. Can., Prog. Abstr., v16, pA3.

Bailes, A.H., Hunt, P.A. and Gordon, T.M. (1991): U-Pb zircon dating of possible synvolcanic plutons in the Flin Flon belt at Snow Lake, Manitoba; *in* Radiogenic age and isotopic studies, Report 4; Geol. Surv. Can., Pap. 90-2, p35-43.

Bell, K. and Macdonald, R. (1982): Geochronological calibration of the Precambrian Shield in Saskatchewan; *in* Summary of Investigations 1982, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 82-4, p17-22.

Bickford, M.E., Collerson, K.D., Lewry, J.F., Van Schmus, W.R., and Chiarenzelli, J. (1990): Proterozoic collisional tectonism in the Trans-Hudson orogen, Saskatchewan; Geol., v18, p14-18.

Byers, A.R. and Dahlstrom, C.D.A. (1954): Geology and mineral deposits of the Amisk-Wildnest Lakes area, Saskatchewan; Sask. Dep. Miner. Resour., Rep. 14., 177p.

Byers, A.R., Kirkland, S.J.T. and Pearson, W.J. (1965): Geology and mineral deposits of the Flin Flon area, Saskatchewan; Sask. Dep. Miner. Resour., Rep. 62, 95p.

Chiarenzelli, J.R., Lewry, J.F. and Landon, M. (1987): Bedrock geology, Iskwatikan Lake area: evidence for Hudsonian juxtaposition of Proterozoic and Archean rocks along a ductile detachment surface; *in* Summary of Investigations 1987, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 87-4, p46-51.

Craig, L. (1989): Geology of the Pelican Narrows area; Unpubl. Ph.D. thesis, Univ. Sask., 268p.

Delaney, G.D., Carr, S.D. and Parrish, R.R. (1988): Two U-Pb zircon ages from eastern Glennie Lake Domain, Trans-Hudson Orogen, Saskatchewan; Geol. Surv. Can., Pap. 88-2, p51-58.

Digel, S. and Gordon, T.M. (1991): Prehnite-pumpellyite to amphibolite facies metamorphism near Flin Flon, Manitoba; *in* Current Research, Pt. C, Geol. Surv. Can., Paper 91-1C, p165-172.

Fedorowich, J., Stauffer, M.R., and Kerrich, R. (in press): Structural setting and fluid characteristics of the Proterozoic Tartan Lake gold deposit, Trans-Hudson Orogen, northern Manitoba; Econ. Geol.

Gordon, T.M., Hunt, P.A., Bailes, A.H. and Syme, E.C. (1990): U-Pb ages from the Flin Flon and Kisseynew belts, Manitoba: Chronology of crust formation at an Early Proterozoic accretionary margin; *in* Lewry, J.F. and Stauffer, M.R. (eds.), The Early Proterozoic Trans-Hudson Orogen, Geol. Assoc. Can., Spec. Pap. 37, p177-200.

Hunt, P.A. and Zwanzig, H.V. (1990): Pre-Missi granitoid domes in the Puffy Lake area, Kisseynew gneiss belt, Manitoba; *in* Radiogenic age and isotopic studies: Report, 3, Geol. Surv. Can., Pap. 89-2, p71-75.

Kober, B. (1987): Single-grain evaporation combined with Pb^+ emitter bedding for $^{207}Pb/^{206}Pb$-age investigations using thermal ion mass spectrometry, and implications for zirconology; Contrib. Mineral. Petrol., v96, p63-71.

Machado, N. (1990): Timing of collisional events in the Trans-Hudson Orogen: Evidence from U-Pb geochronology for the New Quebec Orogen, the Thompson Belt, and the Reindeer Zone (Manitoba and Saskatchewan); *in* Lewry, J.F. and Stauffer, M.R. (eds.), The Early Proterozoic Trans-Hudson Orogen, Geol. Assoc. Can., Spec. Pap. 37, p433-441.

Stauffer, M.R. (1984): Manikewan: an early Proterozoic ocean in central Canada, its igneous history and orogenic closure; Precam. Res., v25, p257-281.

__________ (1990): The Missi Formation: an Aphebian molasse deposit in the Reindeer Lake zone of the Trans-Hudson Orogen, Canada; *in* Lewry, J.F. and Stauffer, M.R. (eds.), The Early Proterozoic Trans-Hudson Orogen, Geol. Assoc. Can., Spec. Pap. 37, p121-142.

Stauffer, M.R. and Mukherjee, A.C. (1971): Superimposed deformations in the Missi metasedimentary rocks near Flin Flon, Manitoba; Can. J. Earth Sci., v8, p217-242.

Stauffer, M.R., Mukherjee, A.C. and Koo, J. (1975): The Amisk Group: an Aphebian(?) island arc deposit; Can. J. Earth Sci., v12, p2021-2035.

Syme, E.C., Hunt, P.A. and Gordon, T.M. (1991): Two U-Pb zircon ages from the western Flin Flon belt, Trans-Hudson orogen, Manitoba; *in* Radiogenic age and isotopic studies: Report 4, Geol. Surv. Can., Pap. 90-2, p25-34.

Van Schmus, W.R., Bickford, M.E., Lewry, J.F. and Macdonald, R. (1987): U-Pb geochronology in the Trans-Hudson Orogen, northern Saskatchewan, Canada; Can. J. Earth Sci., v24, p407-424.

Chemical Signatures of Igneous and Metasedimentary Rocks from the Pelican Slide Area: Implications for Their Sources and Tectonic Environments

M. Sun[1], M.R. Stauffer[1], J.F. Lewry[2], G. Edwards[3], R. Kerrich[1] and T.K. Kyser[1]

Sun, M., Stauffer, M.R., Lewry, J.F., Edwards, G., Kerrich, R. and Kyser, T.K. (1991): Chemical signatures of igneous and metasedimentary rocks from the Pelican Slide area: Implications for their sources and tectonic environments; *in* Summary of Investigations 1991, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 91-4.

The Trans-Hudson Orogen (Hoffman, 1981) is the largest coherently preserved and exposed Early Proterozoic orogenic belt in the world. In northern Saskatchewan and Manitoba, it mainly comprises two contrasting subdivisions, the ensialic Cree Lake Zone and the juvenile Reindeer Zone (Stauffer, 1984). The latter incorporates disparate lithostructural domains dominated by Early Proterozoic arc volcanics, volcanogenic greywacke-pelite assemblages and plutons of various ages. Recent work (Lewry *et al.*, 1990) suggests that most of the exposed Reindeer Zone southeast of the Wathaman batholith is essentially allochthonous, comprising imbricated thrust packages and nappe sheets soled by initially gently dipping ductile high-strain zones. This allochthonous pile may be tectonically emplaced across more-or-less continuous lower plate Archean continental basement along a fundamental basal movement zone.

An exposed part of such a basal sole to the Early Proterozoic allochthons may be represented by the 'Pelican Slide' (Lewry *et al.*, 1989), located in the Hanson Lake block (Figure 1). This feature comprises an early-tectonic, heterogeneous ductile high-strain to mylonitic assemblage of high metamorphic grade orthogneisses and paragneisses which structurally overlie, and envelop the mainly monzocharnockitic Sahli and MacMillan Point Archean 'window' forming the core of the Hanson Lake block (Macdonald, 1974). This largely mylonitic package separates Archean basement from the overlying Early Proterozoic allochthons which form higher structural levels of the Hanson Lake block and the adjoining Kisseynew and Flin Flon domains. Although complexly refolded, the Pelican Slide is thought to have been an initially gently dipping zone, along which regional lateral displacement on the order of tens or possibly hundreds of kilometres may have occurred (Lewry *et al.*, 1989). Thus, the Pelican Slide may be one of the most important structures in this part of the Trans-Hudson Orogen.

The Pelican Slide includes both migmatitic metasedimentary genisses and a broad range of mafic to granodioritic-granitic orthogneisses displaying extremely variable strain intensity and of widely differing age relative to the ductile shear zone fabric. They include pre-kinematic, early- to late synkinematic and post-kinematic major plutons, minor intrusive sheets and anatectic melt fractions (Lewry, 1990).

In June, 1991, 70 samples were collected from the Pelican Slide zone (Figure 1), including metasediments, pre-, syn- and post-tectonic intrusions and anatectic phases. Our objective is to study the geochemistry, pressure-temperature-time and fluid evolution (PTtf) of this zone as part of the Trans-Hudson LITHPROBE project. The first step toward evaluating the PTtf history is detailed study of the petrographic and geochemical character of the rocks involved, initial results of which are reported here.

1. Samples in this Study

a) Archean Inlier

Two samples from the Sahli Granite and one from the McMillan Point Granite were selected. The samples are ~2500 Ma old charnockitic granites (Lewry *et al*, 1987).

b) Proterozoic Intrusions

Proterozoic intrusions include granite, granodiorite, and tonalite, to quartz diorite, and have U-Pb zircon ages between 1870 and 1835 Ma (Lewry *et al*, 1987). The granodiorite and tonalite have strain fabrics increasing in intensity toward the Archean inlier and some are mylonitic; accordingly they have been considered to be early-tectonic plutons. The high-strain pinkish granites intruding the granodiorite are interpreted as syntectonic intrusions, whereas the relatively undeformed rocks of the Jan Lake Granite suite are considered to be post-tectonic. Three granodiorites, one tonalite, and one quartzdiorite representing early-tectonic plutons, two pinkish syntectonic granites, and two Jan Lake granite samples were selected for this study.

---

(1) Department of Geological Sciences, University of Saskatchewan.
(2) Department of Geology, University of Regina.
(3) University of Athabasca, Alberta.

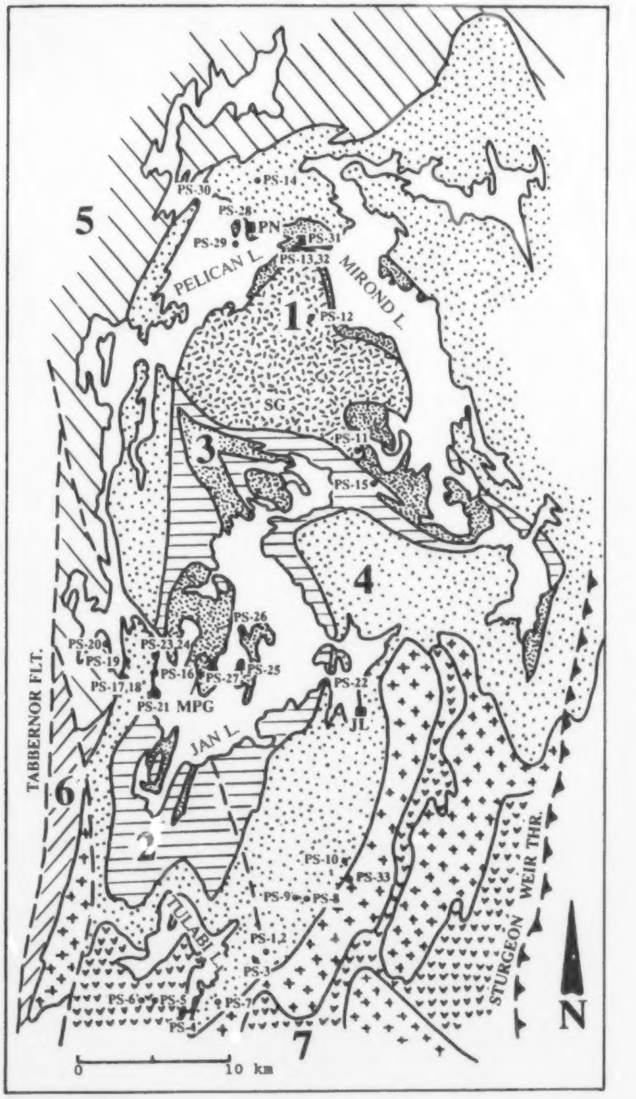


Figure 1 - Simplified geological map of the Jan Lake area, after Lewry et al. (1989), illustrating sample locations. Lithological units as follows: 1-Archean granitoid inliers; 2-Pelitic gneiss assemblage; 3-Quartofeldspathic gneisses; 4-Mixed hornblendic and porphyroclastic gneisses; 5-Kisseynew Gneisses (high-grade pelitic-psmmopelitic); 6-Lower grade metasediments along the Tabbernor Fault zone; 7-Hanson Lake volcanics (low to medium grade) and plutons; PN-Pelican Narrows settlement; JL-Jan Lake resort; SG-Sahli Granite; MPG-McMillian Point Granite.

c) Pegmatites

Pegmatites with different intensities of tectonic fabric are widely distributed in the area. According to their fabric and intrusive relationships, they have been recognized as syn-, late-, and post-tectonic pegmatitic dykes, which are intensely deformed, mildly deformed and undeformed respectively. Four syntectonic, one late-tectonic, and two post-tectonic pegmatites were chosen for chemical study.

d) Mafic Rocks

Mafic rocks in the region include arc volcanics between 1.89 and 1.87 Ga (Rb-Sr dates, Watters and Armstrong, 1985), and mafic intrusions of various dimensions that have ages of 1.88 to 1.79 Ga or younger (Sm-Nd isochron dates, Hegner *et al*, 1989). Two volcanic rocks and five samples from mafic dykes intruded by the Proterozoic granodiorite have been considered as representative of pre- and early-tectonic mafic rocks respectively.

Five samples from mafic dykes intruding the Proterozoic granodiorite are interpreted as syn-tectonic dykes. These dykes are either intruded by late pegmatite or disrupted by tectonic movements.

A (200 to 300 m diameter) pyroxenite body with a round outcrop shape is regarded as post-tectonic; two samples have been analyzed. This body is not deformed and has a chilled margin. The country rocks are cracked around the dyke edges.

e) Metasedimentary Rocks of the Pelican Slide

Eleven samples of metasediments were taken from shoreline exposures along the south part of Pelican Lake (Figure 1). At outcrop scale, the metasediments are generally contorted and injected by variable amounts of leucosome, providing gross evidence for metamorphic conditions sufficient to produce melting (diatexite). Based on mineralogical and textural variations, most of the original sediments were probably mixtures of pelites and psammites. The texture of all samples is granoblastic. The typical metapelitic assemblage is plagioclase + quartz + biotite + garnet + K-feldspar ± cordierite ± sillimanite ± muscovite + accessory minerals.

Sillimanite, invariably coexisting with cordierite, occurs as fine needles in plagioclase, larger isolated euhedral crystals, polycrystalline knots, and fibrolite replacing biotite. In sample PS-21, one anhedral sillimanite grain is overgrown by cordierite. Cordierite is anhedral, and in some samples is partly to completely pinitized. Cordierites analyzed by electron microprobe have molar ratios for Fe/Fe+Mg from 0.34 to 0.40.

The mineral assemblage noted above records peak metamorphism within the upper amphibolite facies. The paucity of prograde muscovite, however, suggests that metamorphism may have reached the amphibolite-granulite transition. The Fe/Fe+Mg ratio of cordierite (0.34 to 0.40) provides a rough upper limit for pressure of 4.5 to 5 kb where cordierite coexists with garnet and biotite. More detailed analysis of the P-T conditions of metamorphism, using garnet-biotite geothermometry and garnet-anorthite and garnet-cordierite geobarometry, is currently in progress.

2. Geochemistry of Igneous Rocks from the Pelican Slide

Major elements were determined by conventional X-ray fluorescence spectrometry, and trace elements by inductively-coupled plasma mass spectrometry (ICP-MS) at the University of Saskatchewan. Reproducibility is generally better than 5 percent for major elements, and 5 to 10 percent for trace elements depending on abundance.

a) Archean Inlier

The Sahli and the McMillan Point Granites have mildly fractionated LREE enriched patterns with $La_n \sim 100$, $Yb_n = 5$ to 20, and $La_n/Yb_n = 6$ to 17 (Figure 2a). This pattern is similar to that of Archean high-Al_2O_3 tonalite-trondhjemite-granodiorite (TTG), e.g. the English River granite (Condie, 1981). The granites are relatively enriched in K, Rb, Ba, Th, and Ce (Figure 3a), and are compositionally similar to granites formed in modern arc environments. They also plot in the Volcanic Arc Granite (VAG) field in Rb vs. (Y+Nb) (Figure 4), Ta vs. Yb, and Rb vs. (Yb+Ta) discrimination diagrams.

b) Early-tectonic Granitoids

Granodiorite

Three granodiorite samples with different strain intensities have REE patterns and La_n/Yb_n ratios (9 to 12) similar to the Archean inlier, but with about half the total REE contents (Figure 2a). The granodiorites are also similar to the Archean granites in trace element distributions (Figure 3a), but contain lower incompatible element contents than the former. They also plot in the VAG field in discriminant diagrams (Figure 4).

Quartz diorite

One quartz diorite is comparable to the Archean granite inlier in terms of REE and trace element distributions (Figures 2a and 3a), and plots in the VAG field in discriminant diagrams (Figure 4).

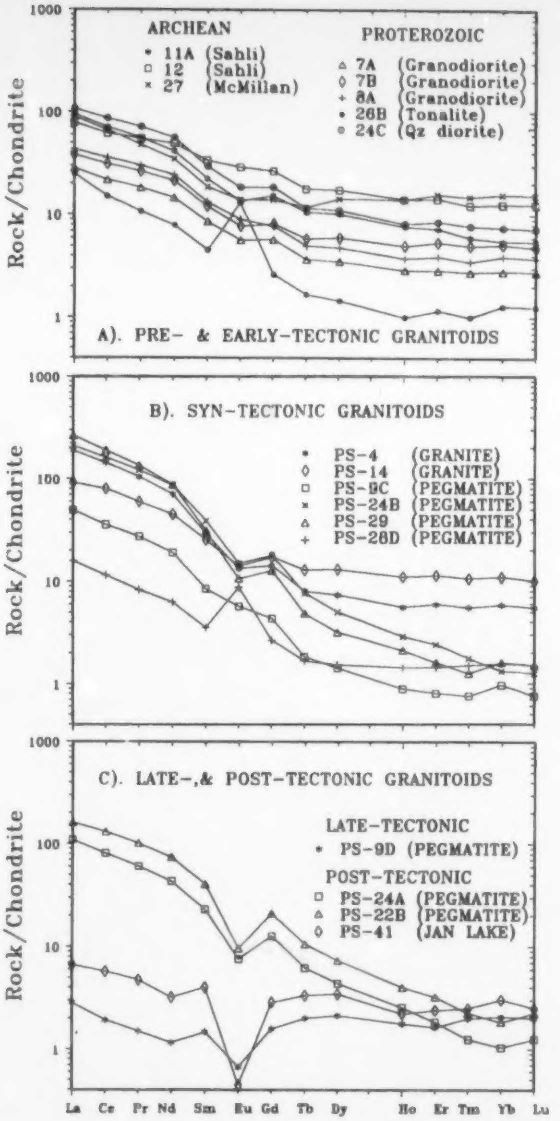


Figure 2 - Chondrite normalized REE distribution of A) Archean and early-, B) syn-, and C) late- and post-tectonic granitoids.

Tonalite

One tonalitic sample has a highly fractionated REE pattern with $La_n/Yb_n = 14.8$, and a pronounced positive Eu anomaly (Figure 2a). The tonalite is enriched in LILE (large ion lithophile elements), but depleted in HFS (high field strength) elements (Figure 3a), and plots in the VAG field (Figure 4).

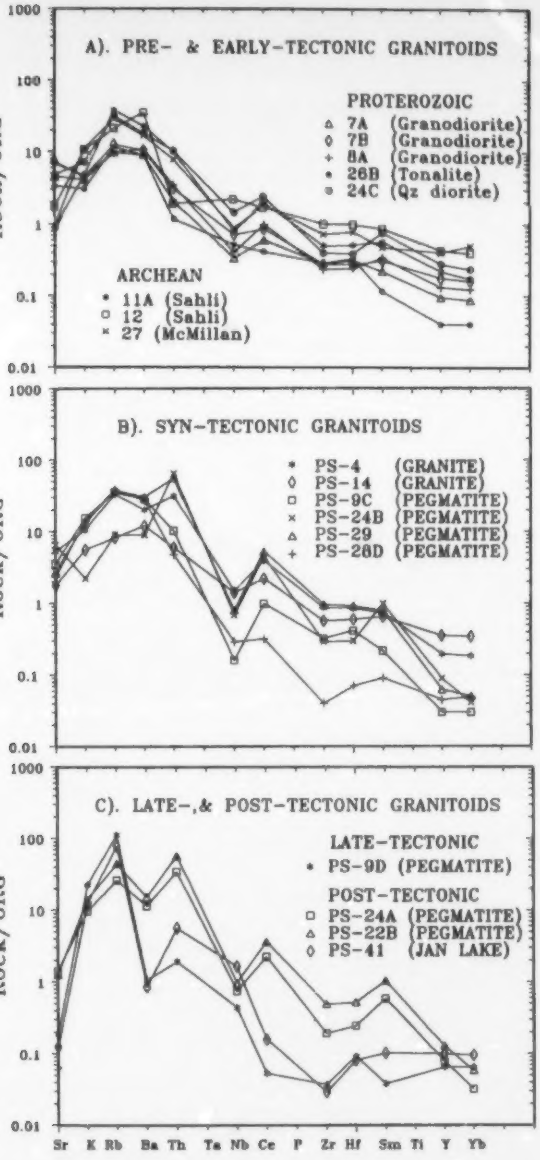


Figure 3 - Ocean ridge granite normalized diagrams for A) Archean and early-, B) syn-, and C) late- and post-tectonic granitoids. Normalizing values from Pearce et al. (1984) and Harris et al., (1986).

c) Syn-tectonic Granitoids

Pink Granite

Two pink granites have REE patterns similar to the Archean inlier, but with a slight negative Eu anomaly; one sample is more LREE enriched, where La_n/Lu_n = 34 (Figure 2b). These rocks are characterized by high Ce and low Nb contents (Figure 3b), and plot in the VAG field (Figure 4).

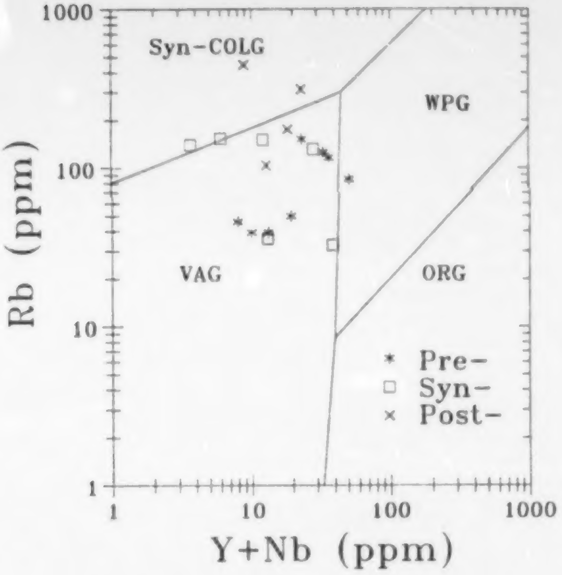


Figure 4 - Granite discrimination diagram (after Pearce et al., 1984) showing pre-, early-, syn-, and late- and post-tectonic granitoids.

Syn-tectonic Pegmatites

Four syn-tectonic pegmatites share highly fractionated REE patterns (La_n/Yb_n = 10 to 180), two have pronounced negative Eu anomalies, one has a positive Eu anomaly, and one no Eu anomaly (Figure 2b). These rocks are generally characterized by high Ce and low Nb, Y, and Yb (Figure 3b). They plot in the VAG field and near the boundary of VAG and syn-collisional granite fields (Figure 4).

d) Late-tectonic Pegmatites

One late-tectonic pegmatite is very depleted in all REE, with a negative Eu anomaly and unfractionated REE ($La_n/Lu_n \approx 1$) (Figure 2c). The late-tectonic pegmatite is characterized by high K, Rb, and Th, low Ba and HFS elements (Figure 3c), and plots in the VAG field (Figure 4).

e) Post-tectonic Granitoids

Post-tectonic Pegmatite

Two post-tectonic pegmatites have fractionated REE distributions where La_n/Yb_n = 72 to 88, and pronounced negative Eu anomalies (Figure 2c). These pegmatites are characterized by high Th, Ce, and Sm, and low Ba, Nb, Zr, Hf, Y and Yb contents (Figure 3c). They plot in the fields of VAG and syn-collisional granites (Figure 4).

Jan Lake Granite

One sample from the Jan Lake Granite is depleted in all the REE, with a pronounced negative Eu anomaly, and

La_n/Yb_n = 3 (Figure 2c). It is enriched in K, Rb, and Th, and depleted in Ba and HFS elements (Figure 3c). It plots in the syn-collisional field (Figure 4).

f) Mafic Rocks

Mafic rocks from the Pelican Slide area are mostly sub-alkaline and tholeiitic. They plot dominantly in the VAB (volcanic arc basalt) field in many discriminant diagrams (e.g. Th vs. Ta). These mafic rocks have flat mildly to LEE enriched REE patterns (Figure 5a, b, and c). There is no significant Eu anomaly. The REE character can be explained either by fractional crystallization of olivine and pyroxene, or by crustal contamination. Some of the pre- and early-tectonic mafic rocks have Ta, Nb, but not Ti, depletions on normalized diagrams; this feature has been interpreted to signify an arc environment.

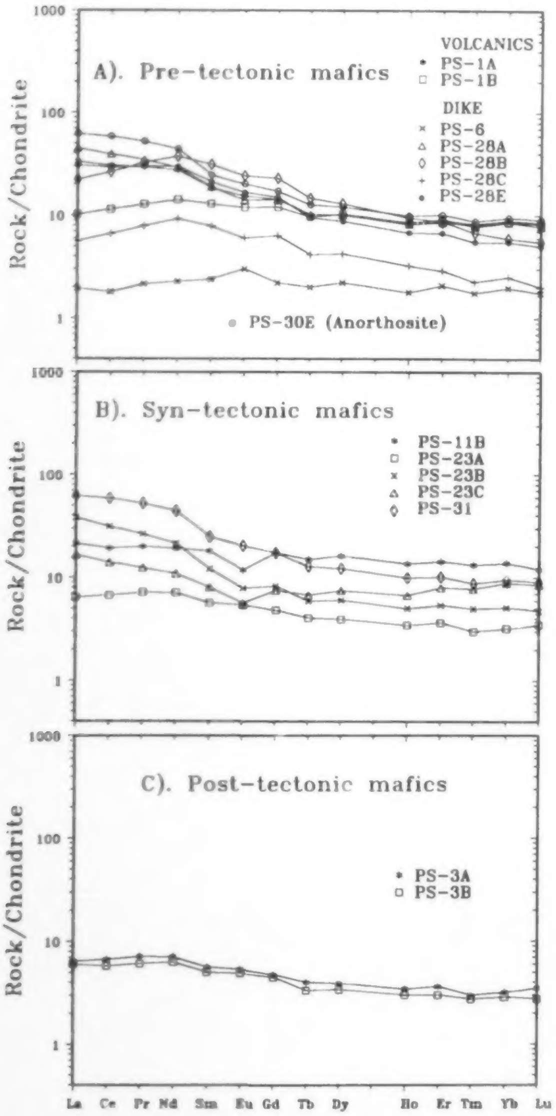


Figure 5 - Chondrite-normalized REE pattern of A) pre- and early-, B) syn-, and C) late- and post-tectonic mafic rocks.

The pre- and early-tectonic mafic rocks have Nb, Zr, and Y depletions, and plot in the field of subduction-related basalts on the Ce/Pb-Nb/Th diagram (Figure 6, Condie, 1990). The La-La/Nb diagram (Jahn, 1990) shows a significant involvement of Archean continental material in the formation of the pre- and early-tectonic mafic rocks (Figure 7). We infer that they were formed in a volcanic arc-like environment proximal to an Archean continent.

The syn-tectonic mafic rocks are not depleted in Nb, but are enriched in Th, and depleted in Zr and Ti. They mainly plot off defined fields in discriminant diagrams (e.g. Figure 6). This could be due to element mobility

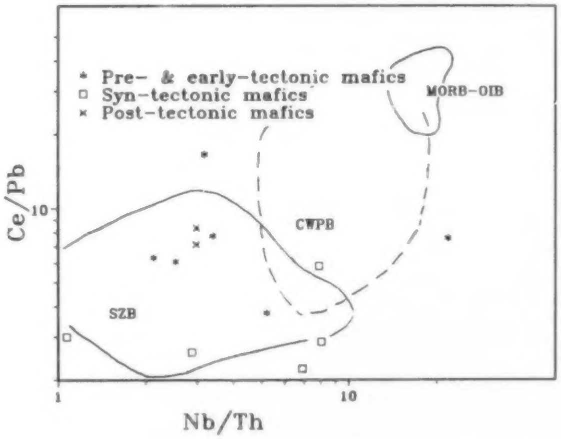


Figure 6 - Ce/Pb vs. Nb/Th diagram (after Condie, 1990) showing pre- and early-, syn-, and late- and post-tectonic mafic rocks.

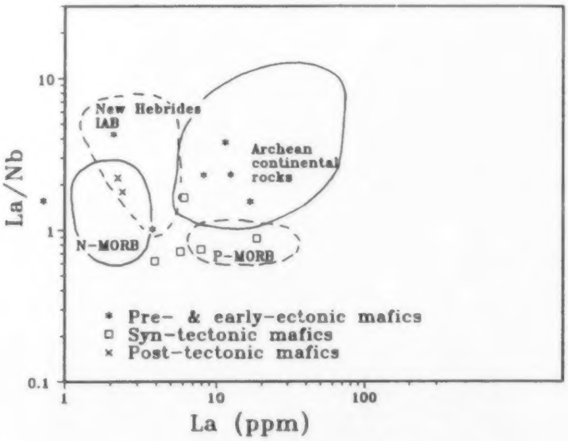


Figure 7 - La/Nb vs. La diagram (after Jahn, 1990) showing pre- and early-, syn, and late- and post-tectonic mafic rocks.

during metamorphism. It appears that they were less contaminated by Archean continental material during their formation (Figure 7).

Samples from the post-tectonic mafic intrusions have flat depleted REE patterns (Figure 5c) and a trace element distribution similar to oceanic island arcs. They also plot in the field for modern IAB, away from the Archean continent field (Figure 7).

g) Summary

Early and syn-tectonic granitic rocks from the Pelican Slide area have a VAG (volcanic arc granite) character, whereas late-stage pegmatites evolved towards the syn-collisional granite field. The Archean granites are LREE and K, Rb, Ba, Th enriched, as is the early-tectonic quartz diorite. Granodiorites are compositionally similar, but feature lower incompatible element contents. An early-tectonic tonalite, however, has a positive Eu anomaly and depletion in HFS elements.

The syn-tectonic pink granites and pegmatites are more REE fractionated, and have high Th, Ce, and low Nb contents. A late-tectonic pegmatite is depleted in all REE and HFS elements, has a negative Eu anomaly, and Th enrichment. Post-tectonic pegmatites are characterized by high Rb, Th, Ce, Sm and low Nb. The post-tectonic Jan Lake Granite is depleted in Ba, REE and HSF elements, and enriched in K and Rb relative to all other granites.

Mafic rocks from the Pelican Slide area are all subduction-related.

3. Discussion

This preliminary geochemical work has shown significant differences in pre-, early-, syn-, late- and post-tectonic granitoids, providing a solid basis for further study of the pressure, temperature, time, and fluid evolution. It is not clear whether the Sahli and McMillan Point Granites are autochthonous or para-autochthonous Archean continental basement, or alternatively whether the granites are part of an allochthonous basement thrust slice transported within the high-strain sequence. The chemical signatures of granodiorite and quartz diorite are similar to the Archean granites, and the pre-, and early-tectonic mafic rocks may have a significant involvement of Archean material. This may lend support for the allochthonous explanation for the Archean inlier. Final resolution of this question awaits further detailed studies.

4. Future Objectives

Further work will address the pressure, temperature, time, and fluid evolution of the Pelican Slide using appropriate geobarometers, geothermometers and geochronology. Nd isotopic data will help to reveal the sources of rocks, and model magma genesis.

5. Acknowledgments

This study was funded by a Trans-Hudson LITHOPROBE grant, and NSERC Infrastructure grant to R. Kerrich and T. K. Kyser. We thank J. Jain for assistance with ICP-MS.

6. References

Condie, K.C. (1981): Archean Greenstone Belts; Elsevier, 381pp.

__________, (1990): Geochemical characteristics of Precambrian basaltic greenstones; *in* Hall, R.P. and Hughes, D.J. (eds.), Early Precambrian Basic Magmatism, Blackie and Son, Glasgow, p40-55.

Harris, N.B.W., Pearce, J.A. and Tindle, A.G. (1986): Geochemical characteristics of collision-zone magmatism; *in* Coward, M.P. and Ries, A.C. (eds.), Collision Tectonics, Geol. Soc., Spec. Publ. no. 19, p67-81.

Hegner, E., Kyser, T.K. and Hulbert, L. (1989): Nd, Sr, and O isotopic constraints on the petrogenesis of mafic intrusions in the Proterozoic Trans-Hudson Orogen of central Canada; Can. J. Earth Sci., v26, p1027-1035.

Hoffman, P.F. (1981): Autopsy of Athapuscow Aulacogen: a failed arm affected by three collisions; *in* Cambell, F.H.A. (ed.), Proterozoic Basins of Canada, Geol. Surv. Can., Pap. 81-10, p97-102.

Jahn, B.M. (1990): Early Precambrian basic rocks of China; *in* Hall, R.P. and Hughes, D.J. (eds.), Early Precambrian Basic Magmatism, Blackie and Son, Glasgow, p294-316.

Lewry, J.F. (1990): Bedrock geology, Tulabi-Church Lake area: derivation and significance of porphyroclastic gneisses in the Pelican Window; *in* Summary of Investigations 1990, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 90-4, p36-43.

Lewry, J.F., Macdonald, R., Livesey, C., Meyer, M., Van Schmus, R. and Bickford, M.E. (1987): U-Pb geochronology of accreted terranes in the Trans-Hudson Orogen, northern Saskatchewan, Canada; *in* Pharaoh, T.C., Beckinsale, R.D. and Rickard, D. (eds.), Geochemistry and Mineralization of Proterozoic Volcanic Suites, Geol. Soc. Spec. Publ. no. 33, p147-166.

Lewry, J.F., Macdonald, R. and Stauffer, M.R. (1989): The development of highly strained rocks in the Pelican Window during high-grade metamorphism and pervasive anatexis; *in* Summary of Investigations 1989, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 89-4, p58-65.

Lewry, J.F., Thomas, D.J., Macdonald, R. and Chiarenzell, J. (1990): Structural relations in accreted terranes of the Trans-Hudson Orogen, Saskatchewan: telescoping in a collisional regime?, *in* Lewry, J.F. and Stauffer, M.R., (eds.), The Early Proterozoic Trans-Hudson Orogen of North America; Geol. Assoc. Can., Spec. Pap. 37, p75-94.

Macdonald, R. (1974): Pelican Narrows (west) area: reconnaissance geological survey of 63M-2(W); in Summary Report of Field Investigations by the Saskatchewan Geological Survey, Sask. Dep. Miner. Resour., p30-37.

Pearce, J.A., Harris, N.B.W. and Tindle, A.G. (1984): Trace element discrimination diagrams for the tectonic interpretation of granitic rocks; J. Petrol., v25, p956-983.

Stauffer, M.R. (1984): Manikewan: an Early Proterozoic ocean in central Canada, its igneous history and orogenic closure; Precam. Res., v25, p257-281.

Watters, B.R. and Armstrong, R.L. (1985): Rb-Sr study of metavolcanic rocks from the La Ronge and Flin Flon Domains, northern Saskatchewan; Can. J. Earth Sci., v22, p452-324.

Origin and Kinematic History of Highly Strained Gneisses in the Jan Lake East Area

Shi Rupan[1] *and J.F. Lewry*[1]

Shi, R. and Lewry, J.F. (1991): Origin and kinematic history of highly strained gneisses in the Jan Lake East area; *in* Summary of Investigations 1991, Saskatchewan Geological Survey, Saskatchewan Energy Mines, Misc. Rep. 91-4.

Previous work has defined a broad zone of heterogeneously highly strained to mylonitic orthogneisses and paragneisses enveloping and overlying the Sahli and MacMillan Point Archean basement inliers in the 'Pelican Window' of the Hanson Lake Block (Lewry and Macdonald, 1988; Lewry *et al.*, 1989; Lewry, 1990). The extent and degree of continuity, however, of this high-strain package along the eastern side of the Hanson Lake dome was poorly defined. This summer, the first author conducted 1:20 000 scale remapping around the eastern part of Jan Lake (Figure 1), an area believed to include eastern extensions of the high-strain assemblage. All or parts of the area were previously mapped by Pyke (1966), Macdonald and MacQuarrie (1978), MacQuarrie (1979) and Lewry *et al.* (1989). Work reported here will form the basis of an M.Sc. thesis, the main objectives of which are to:

a) more clearly define the distribution and nature of ductile high-strain rocks formed during early deformation events in the region,

b) document strain gradients across the high-strain gneisses and their less-strained protoliths, and establish sense of movement in the high-strain zone, and

c) identify the protoliths of the varied lithological components of the high-strain package, though field-structural, petrographic and geochemical investigations.

1. Rock Units

a) Unit 1: Pelite Metatexite-diatexite

Migmatitic gneisses derived from dominantly pelitic to psammopelitic sediments extend along the eastern shore of Jan Lake (Figure 1). They range from metatexites comprising dominant schistose to gneissose pelitic to psammopelitic paleosome with subordinate leucogranitic-pegmatitic neosome, to diatexites comprising 70 percent or more granitoid leucosome and plentiful ultramafic melanosomal 'restite', but retaining little or no clearly defined paleosome.

Paleosome Components

Metasedimentary paleosome mainly comprises quartz, plagioclase (quartz generally more abundant than plagioclase), subordinate K-feldspar and biotite content ranging from about 10 to 30 percent. Dominant pelitic gneisses typically contain garnet and sillimanite which are typically most abundant in the melanosome. Cordierite, although not obvious in hand specimen, is seen in thin section (Abbas-Hasanie and Lewry, this volume). Garnet crystals range from a few millimetres to several centimetres (locally to 7 cm) across. Generally, they form apparently slightly deformed porphyroblasts, although highly strained, pre- to syn-tectonic ovoidal garnet porphyroclasts with prominent recrystallization tails occur locally. Garnet generally forms less than 5 percent of the rock but locally ranges to 15 percent. Boudinaged garnet-rich (up to 40 percent) bands 0.5 to 2.0 m wide, parallel to the main foliation, occur in a few outcrops.

Relict primary layering is locally well defined by medium- to fine-grained, mid- to dark-grey, relatively quartz-rich psammitic to psammopelitic metagreywacke beds, typically 3 to 10 cm but locally up to 1 m or more thick. Psammitic to psammopelitic members are also dominant locally. Psammitic layers typically contain only minor neosomal segregations, and are sparsely garnetiferous.

Mafic amphibolite bands, comprising medium- to coarse-grained hornblende and plagioclase, ranging in thickness from centimetres to several metres, are typically boudinaged, locally rotated, and folded. These rocks are interpreted as mafic dykes rather than supracrustal layers (Lewry *et al.*, 1989).

Neosome components

The pink leucogranitic to pegmatitic leucosome varies from discontinuous to semi-continuous '*lits*' and lenses parallel to paleosome foliation ('stromatic' migmatites) to more irregular, partially discordant and highly folded masses enclosing subordinate paleosomal schlieren, and also forms isolated minor fold noses and irregular patches ranging from a few millimetres to over a metre across. Quartz, plagioclase and K-feldspar with minor biotite and, locally, cordierite and sillimanite are the main mineral components. Fabric relations between

---

(1) Department of Geology, University of Regina, Regina, Saskatchewan.

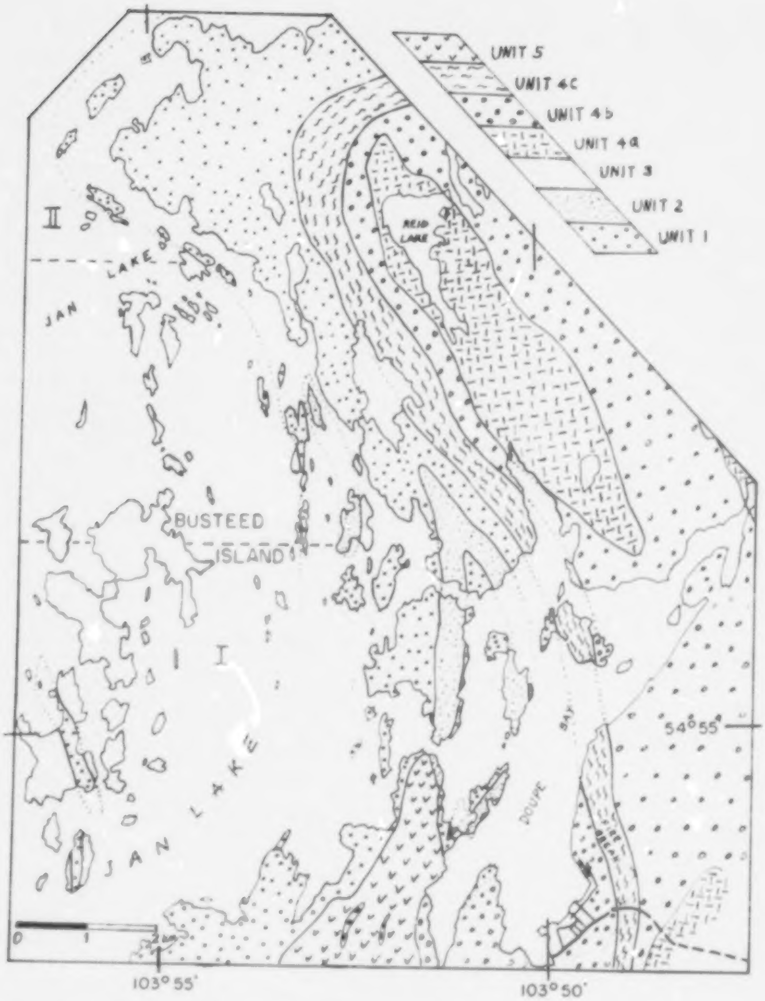


Figure 1 - Geological sketch map of the eastern Jan Lake area. Unit 1, pelite metatexite-diatexite; Unit 2, megacrystic biotite tonalite; Unit 3, non-megacrystic tonalite-granodiorite; Unit 4a, less strained hornblende-biotite granodiorite complex; Unit 4b, moderately to highly strained porphyroclastic granodioritic gneisses; Unit 4c, protomylonitic-mylonitic porphyroclastic, 'beaded' gneisses; Unit 5, Jan Lake granite.

leucosomal *lits*, foliation and early minor folds indicate multiphase leucosome development or, more likely, continuous leucosome generation throughout early deformational events. Very early leucosome layers parallel the main foliation and are commonly isoclinally folded, ribboned and disrupted. Later leucosome components are variably concordant and discordant to the main foliation and are either apparently undeformed or are folded only by late minor folds.

Commonly, the leucosome bodies are prominently bordered by biotite-rich, garnet- or garnet-sillimanite-bearing melanosomal selvages up to several centimetres thick. Elsewhere, particularly in the diatexites, the melanosome forms plentiful lensoid to irregular patches external to or within the leucosome.

Fabric

The anatectic metasediments are strongly foliated, gneissic and schistose, and locally intensely laminated. Foliation is accentuated by highly elongate quartz and quartz-feldspar lenses which probably represent disrupted early neosome *lits*. Pre- and early synkinematic lithological elements, such as primary psammitic-psammopelitic layers, mafic dykes and early neosome, are completely transposed into the main foliation, disrupted, boudinaged, folded and rotated. Strongly laminated variants generally contain a variable amount of ovoidal to rounded white plagioclase and/or pink K-feldspar porphyroclasts and smaller 'beads', generally from a few millimetres to about 2 cm across, but locally up to 5 cm. Some units, typified by a very biotite-rich matrix enclosing abundant ovoidal to rounded plagioclase porphyroclasts, resemble the biotite-rich porphyroclastic gneisses of the Church Lake area (Lewry, 1990) which were interpreted as 'mylonitic' derivatives of pelite diatexite.

b) Unit 2: Megacrystic Biotite Leucotonalite-granodiorite

Megacrystic biotite leucotonalite-granodiorite occurs on many islands in the main basin of Jan Lake, and is intercalated with pelite diatexites along the eastern shore of Jan Lake and on islands of western Doupe Bay. The main rock types are all light grey to grey-white in fresh and weathered surfaces but vary from medium-coarse to very fine grained. White plagioclase, quartz (feldspar generally more abundant than quartz), variable amounts of biotite, ranging from less than 5 percent up to 15 percent locally, and local minor garnet are the mineral components. K-feldspar generally seems to be a subordinate constituent. The unit commonly incorporates up to 20 percent or more pink granitic neosome, typically as transposed and strongly foliated layers and thin laminae concordant to kinematic foliation.

Strain effects are extremely variable, both regionally and at outcrop scale. The least strained variants are relatively homogeneous, weakly foliated rocks which typically contain up to 10 percent subidioblastic white plagioclase megacrysts. The only other inhomogeneity in such rocks is variably deformed neosomal granitic material. Small ductile high-strain zones parallel to otherwise generally weak foliation are common. More highly strained types are medium to fine grained, intensely foliated and laminated, and locally include isolated, ovoidal to rounded plagioclase and/or K-feldspar porphyroclasts about 0.5 to 1.5 cm across. In some places, they also contain rotated mafic boudins. In many places these rocks contain pinkish-grey, medium-grained, quartzofeldspathic layers with sparse, ovoidal K-feldspar porphyroclasts about 1 cm across as well as large and highly ribboned quartz lenses and small disrupted pegmatite sheets. Such layers are generally parallel or slightly oblique to the main foliation and are interpreted as highly strained intrusive or neosomal felsic sheets rather than primary layers. At one locality, a pelite diatexite block 1.5 m across, occurring within moderately, highly-strained, biotite tonalitic gneiss appears to be a xenolith.

On the basis of mineral content, lithological homogeneity, evident strain gradation between coarse-grained megacrystic variants and finer-grained types, a xenolithic inclusion, and lack of layering other than that resulting from transposition of neosome components, this unit is thought to be derived from plutonic protoliths.

c) Unit 3: Non-megacrystic Tonalite to Granodiorite

Non-megacrystic tonalite to granodiorite, exposed mainly on Busteed Island and small islands to the north and south, formerly identified as possible psammitic-psammopelitic metasediments by Macdonald and Macquarrie (1978), was re-interpreted as highly-strained plutonic rocks, at least in part, by Lewry *et al.* (1989). Generally, their paleosome is uniformly medium to fine grained, incorporating a variable proportion of continuous to discontinuous, coarser-grained leucosomal quartzofeldspathic layers, patches and lenses. These are commonly bordered by biotite-rich melanosomal selvages, suggesting that the leucosome represents *in situ* anatexis, at least in part. The neosome locally contains small dispersed garnets or garnet aggregates.

As in unit 2, these rocks are heterogeneously strained. Locally, for example at the north end of Busteed Island, relict, weakly foliated and homogeneous medium-grained tonalite-granodiorite with plutonic fabrics is preserved. Commonly observed strain gradation suggests that more strongly foliated, medium- to fine-grained variants are derived from similar plutonic protoliths via progressive ductile high-strain, dynamic recrystallization, and accompanying shredding and transposition of neosome. Quartz in the latter is intensely ribboned and large feldspars locally form ovoidal to rounded porphyroclasts averaging 0.2 to 0.5 cm across, some with well-developed tail complexes. In places, the fine-grained matrix includes sparse small, white to pink, rounded feldspar beads which may represent comminuted primary igneous crystals or derive from disruption of neosomal veins.

Outcrop areas affected by heterogeneous intermediate strain commonly show narrow ductile mylonite zones subparallel to foliation in less strained parts of the outcrop. Strain gradients are clearly seen across the margins of these zones. In some localities, as for example in a few outcrops examined outside the main map area along the southwest shore of Mirond Lake, lensoid 'tectonic fish' of coarser-grained tonalite are preserved in generally fine-grained, intensely laminate-foliated quartzofeldspathic gneisses. Foliation in the more highly strained components is locally defined by extreme quartz ribboning and more generally by development of a fine tectonic lamination. Minor high-strain zones, some displaying clear C-S fabrics, are also locally developed along the contacts between neosomal pegmatite sheets and tonalitic host rocks. These are generally subparallel to the main foliation but may represent a later phase of deformation.

Locally, unit 3 rocks enclose blocks of medium- to fine-grained, garnetiferous hornblende gneiss with well-developed mafic to felsic layering. These blocks are interpreted as metavolcanic xenoliths. Isolated to semicontinuous inclusions of uniform, medium- to coarse-grained L-tectonite amphibolite are interpreted as boudinaged and disrupted mafic dykes.

Subunit 3a

Subunit 3a, occurring mainly on Busteed Island and small islands to the north, is the dominant rock type. It is medium to fine grained, light grey to pale buff on weathered surfaces and pinkish grey or pink on fresh surface. The paleosome groundmass mainly comprises white and pink feldspar (55 to 60 percent), quartz (20 to 40 percent) and biotite (5 to 10 percent). Discrete white to pink neosomal quartzofeldspathic bands, ranging from a few millimetres up to 3 or 4 cm thick, generally form 10 to 15 percent of the rock but locally up to 40 percent. This subunit is generally thought to be derived from plutonic protoliths.

Subunit 3b

Subunit 3b, mainly occurring on the islands south of Busteed Island and in the east bay of Jan Lake, has a high proportion of pink granitoid neosome, in continuous to semicontinuous layers, elongate lenses, irregular patches and ill-defined 'sweats', typically forming 30 to 50 percent but locally up to 80 percent of outcrop areas. The intervening medium- to fine-grained quartzofeldspathic host rock contains up to 40 percent biotite and may be either highly modified melanocratic paleosome or restitic melanosome. In places, particularly biotite-rich schlieren tend to be concentrated along the margins of leucosome masses. Tight to isoclinal intrafolial minor folding of the neosome is common. Unlike the pelite diatexites of unit 1, this subunit contains little or no garnet or sillimanite; nonetheless, the high biotite content suggests that this subunit may derive from a metasedimentary protolith.

Subunit 3c

Subunit 3c, occurring mainly on Busteed Island, typically has a well developed gneissic foliation and mineral lineation defined by individual crystal alignment, multicrystal quartz ribbons and multicrystal monomineralic aggregates of feldspar and biotite, suggestive of an originally coarse grain-size. Generally, they contain about 10 to 20 percent biotite (intermediate to that in subunits 3a and 3b), 50 to 60 percent feldspar and about 30 percent quartz. Semicontinuous pink quartzofeldspathic neosome layers averaging 1 to 3 cm in thickness lie parallel to the main foliation. In places, the rocks have a strong tectonic lamination accompanied by prominent quartz ribboning and small isolated rounded feldspar porphyroclasts 0.2 to 0.4 cm across. Large pink feldspar porphyroclasts are dynamically recrystallised into medium- to fine-grained crystal aggregates. We interpret this subunit to be derived from a relatively coarse-grained granitic-granodioritic plutonic protolith.

As in Unit 2, this subunit locally has grey, medium- to fine-grained, quartzofeldspathic layers ranging from 1.5 to 15 cm thick, parallel or subparallel to foliation and containing prominent large, quartz ribbons and small, sparse ovoidal feldspar porphyroclasts. The grey layers are locally bordered or injected by small undeformed pegmatite sheets. On the basis of field relations, they are interpreted as syn-tectonic felsic dykes, injected along small shear zones, with late pegmatite sheets injected along the contact zones.

d) Unit 4: Variably Strained Hornblende-Biotite Granodiorite

Variably strained hornblende-biotite granodiorite underlies most of the area around and to the north and east of Doupe Bay. This unit is broadly granodioritic in composition, mid- to light-grey, and weakly to intensely foliated. Hornblende and biotite occur in variable amounts. At outcrop scale, much of the unit, particularly the more highly strained variants, incorporates relatively coarser and apparently less deformed lenses and irregular patches of granodioritic neosome. The neosome is characterized by prominent hornblende blastesis and is locally dominant. Boundaries between neosomal lenses and more strongly foliated finer grained paleosome are usually ill-defined and gradational. Locally, the neosomal zones include irregular to lensoid, coarse-grained ultramafic hornblende aggregates with sharp to diffuse boundaries; these are mostly interpreted as melanosomal segregations but some may represent disrupted mafic inclusions. Ovoidal to rounded plagioclase feldspar porphyroclasts and small 'beads', typically ranging in size from 0.1 to 1.2 cm across, are a prominent feature of the paleosome.

The granodiorite complex displays a clear, albeit irregular increase in ductile strain from east to west across the Doupe Bay area, from unequivocally plutonic, weakly to moderately foliated granodiorite through increasingly highly strained and finer-grained granodioritic gneisses, to generally fine-grained protomylonitic and mylonitic gneisses. Boundaries between these strain states are irregularly gradational, heterogeneous and apparently anastomosing.

Increase in strain is defined by: a) general decrease in grain size, b) more prominent ribboning of quartz grains, until advanced stages of strain where they disappear into the groundmass, c) increasing 'beading' of groundmass plagioclase grains, d) eventual replacement of a well-foliated gneissic groundmass by a less clearly foliated medium- to fine-grained dynamic recrystallization matrix, e) progressive elongation, disruption and rotation of white quartzofeldspathic neosomal layers and pink pegmatite sheets, and f) progressive isolation and rounding of large K-feldspar porphyroclasts which either originated as discrete porphyroblasts or derive from disruption of thin pegmatite neosome layers.

On the basis of systematic variation in such strain indicators, the complex is divided into three subunits.

Subunit 4a

Subunit 4a comprises weakly to moderately strained, coarse- to medium-grained hornblende-biotite granodiorite with variable amounts of plagioclase megacrysts which locally display tight intrafolial folds defined by early neosomal quartzofeldspathic layers. Neosomal leucogranodiorite layers and patches are also locally deformed together with the main foliation by a later generation of open to tight minor folds. Weak quartz ribboning and weakly elongated to rounded plagioclase megacrysts are seen locally.

Subunit 4b

Subunit 4b includes moderately highly strained to protomylonitic, porphyroclastic granodioritic gneisses which constitute a markedly heterogeneous ductile strain transition between subunits 4a and 4c. In general, these rocks are medium grained and strongly foliated, with a variable percentage of ovoidal to rounded plagioclase porphyroclasts, 'beaded' groundmass plagioclase, and large ovoidal to rounded K-feldspar porphyroclasts, locally with medium- to fine-grained dynamic recrystallization tail complexes. Some semi-isolated large plagioclase and K-feldspar porphyroclasts are wrapped about by large intensely ribboned quartz subdomains and are probably derived from shredded neosomal pegmatite veins.

Subunit 4c

Subunit 4c comprises protomylonitic to mylonitic 'beaded grey gneisses' which constitute a relatively well-defined ductile high-strain zone. In the north, this zone lies between unit 1 pelite diatexites and less strained subunits of the granodiorite complex. In the south, it lies entirely within unit 4. The subunit typically displays variable amounts of small discrete grey-white feldspar 'beads' in a medium- to fine-grained dynamic recrystallization matrix. In addition, highly disrupted neosomal pegmatite sheets and small to large ovoidal to rounded plagioclase and K-feldspar porphyroclasts with or without tails also occur in places. Locally they are intensely laminated, with only sparse sporadic small feldspar 'beads'. Commonly, the high-strain fabric is obscured by extensive hornblende blastesis; the hornblende crystals are generally idioblastic, 0.4 to 1.5 cm in size and clearly post-date the high-strain fabric, a feature also noted in the ductile shear zone in the Tulabi Lake area (Lewry, 1990).

e) Unit 5: Jan Lake Granite and Pegmatite

Late- to post-tectonic granites and pegmatites of the Jan Lake suite are dominant in the Doupe Bay area. Generally, they form medium- to very coarse-grained, massive to weakly foliated, cross-cutting to subcondordant sheets, dykes, apophyses and larger irregular intrusive bodies. Commonly, foliation in Jan Lake granite bodies is most prominent near contacts with older country rocks. Pegmatite sheets cut all other rocks in the area and commonly display small brittle to ductile shear zones, either at their contacts with surrounding rocks or in the middle of the sheets. Their emplacement was possibly controlled, in part, by late-stage brittle-ductile shear zones.

An east-west-trending, uniform, coarse- to medium-grained, biotite-rich granitic dyke cutting unit 3 rocks on Busteed Island is included in unit 5 because of its generally weak foliation.

Jan Lake granite and pegmatite are the dominant rock types in the Doupe Bay area, with rocks of unit 4 occurring mainly as highly subordinate, large and small xenoliths and rafts within them. However, in order to emphasize the structurally more important strain distribution in rocks of unit 4, the extent of the Jan Lake suite is not fully shown on the accompanying map.

2. Structure

All rocks in the area other than those in unit 5 underwent polyphase ductile deformation. Many outcrops show complex minor fold interference patterns, defined mainly by leucosomal *lits*, mafic folia and prior foliation.

a) Early Kinematic Fabrics

Details of early structural sequence are best preserved in unit 1 pelite diatexites. Here, an S_1 gneissic foliation, paralleled by primary layering, early neosomal *lits* and mafic folia, is locally distinguishable. In general, however, it is completely transposed into, and indistinguishable from a dominant regionally developed S_2 foliation which is axial planar to tight to isoclinal and commonly isolated intrafolial F_2 minor folds defined by transposed primary layering and early leucosomal-melanosomal laminae (Lewry et. *al.*, 1989). Such folds are typified by highly attenuated to discontinuous limbs and thick, 'acute' hinge zones. Fold amplitudes range from a few centimetres to several tens of centimetres. Such intrafolial folds are less common in other rock units. In rare cases, minor intrafolial isoclines are defined by primary psammitic layers (S_0) paralleled by finely foliated and disrupted quartzofeldspathic leucosome layers (S_1) incorporating ovoidal feldspar porphyroclasts. These S_1 fabric elements are deformed around the fold noses, where they are oblique to dominant S_2 axial plane fabrics. Even more rarely, F_2 folds are seen to refold earlier F_1 intrafolial isoclines defined by primary lamination. In other rock units, no distinction between D_1 and D_2 planar fabrics and minor folds was established.

The extreme regional and local ductile strain variation is clearly related to the D_1/D_2 progressive strain continuum, since mylonitic foliation documented in many rock units is clearly refolded by D_3 and D_4 folds. S_1/S_2 strain gradation from plutonic protoliths to essentially mylonitic and locally porphyroclastic quartzofeldspathic gneisses is well documented in units 2 and 3. Both local and broader regional gradation from relatively weakly foliated granodiorites to their protomylonitic and mylonitic derivatives in unit 4c is also well-defined. The main high-strain zone represented by unit 4c appears to be a northerly extension of a comparable ductile shear zone in granodioritic to tonalitic rocks identified by Lewry (1990) farther south, in the northeastern part of the Tulabi-Church Lakes area. Its divergence from the pelite diatexite-granodiorite boundary suggests that it may represent a higher level splay rooting in a more fundamental movement zone within the highly incompetent pelite diatexites. Although strain variation in the latter was not consistently identified; its lithological character and the evident presence of a high proportion of anatectic melt throughout D_1/D_2 deformation suggests that it is likely to have suffered even greater strain than the plutonic units.

The most highly deformed components of units 1 to 4 all locally display evidence of a significant simple shear strain component , including prominent rotated (delta-type) tailed feldspar porphyroclasts, rotated mafic boudins and more isolated mafic inclusions, and local C-S fabrics. Such shear-sense indicators observed on subhorizontal outcrop surfaces of unit 1 pelite diatexite overwhelmingly indicate dextral shear-sense, whereas indicators in subunits 4b and 4c mostly show sinistral shear-sense. Stretching lineation related to these indicators was only rarely identified and the three-dimensional significance of the variation in shear-sense is presently unclear.

b) Late Fold Phases

The S_1/S_2 ductile strain fabric, including mylonitic variants thereof, is widely refolded by open to tight F_3 minor folds which are locally accompanied by an S_3 axial plane fabric oblique to early foliation. A later F_4 open fold set can also be differentiated locally.

Most open to tight folds (Figure 2) are thought to be F_3 structures. Their plunge variation is probably the result of both variation in prior attitude of S_1/S_2 foliation and broad F_4 regional refolding. In the south, the minor folds generally plunge towards the south-southwest approximately coaxial with major F_3 regional folds farther south (Lewry *et al.*, 1989). Early foliation in the northern part of the area defines a major open fold with a statistical plunge of 40° toward 110°. This orientation is discordant with the general northeasterly trend of F_4 regional folds (Lewry *et al.*, 1989) and the fold is interpreted as an F_3 structure.

3. Metamorphism

Coexisting garnet, sillimanite and K-feldspar, and general absence of muscovite in diatexitic pelites, high

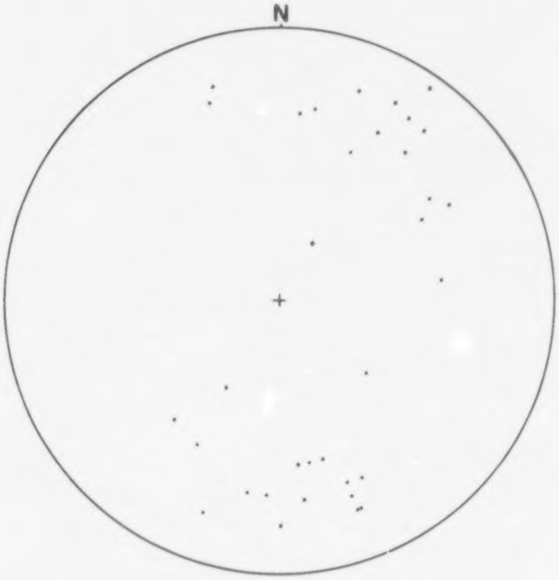


Figure 2 - Stereoplot of hingelines of post-D_2 minor tight to open folds (34 points)

percentage partial melting of both pelitic and plutonic protoliths and deformational relationships of neosomal components indicate that high grade (upper amphibolite facies) conditions were attained and probably maintained throughout the D_1 to D_2 events. Anatexis appears to have occurred both prior to and during development of the high-strain protomylonitic to mylonitic fabric in the orthogneiss units.

4. Preliminary Conclusions

1) Most or all of the rocks included in units 2, 3 and 4 in the eastern Jan Lake area are derived from plutonic protoliths, rather than from psammites and crystal tuffs. Their present character is the result of heterogeneously high ductile strain related to early phases of regional deformation.

2) Most rocks in the area have undergone at least two phases of partial melting, four phases of fold deformation and variable simple shear deformation. Partial melting took place prior to D_3 folding, and prior to and/or during high-strain simple shear deformation.

3) Most of the map area attained middle to upper (probably uppermost) amphibolite facies conditions during the D_1 to D_2 folding and essentially coeval ductile, high-shear-strain deformation.

5. References

Lewry, J.F. (1990): Bedrock geology, Tulabi-Church Lakes area: derivation and significance of porphyroclastic gneisses in the Pelican Window; *in* Summary of Investigations 1990, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 90-4, p36-43.

Lewry, J.F. and Macdonald, R. (1988): Observations on deformation in the Glennie Domain and Hanson Lake Block; *in* Summary of Investigations 1988, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 88-4, p35-41.

Lewry, J.F., Macdonald, R. and Stauffer, M.R. (1989): The development of highly strained rocks in the Pelican Window during high-grade metamorphism and pervasive anatexis; *in* Summary of Investigations 1989, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 89-4, p58-65.

Macdonald, R. and MacQuarrie, R.R. (1978): Geological re-investigation mapping, Jan Lake area (part of NTS area 63M); *in* Summary of Investigations 1978, Saskatchewan Geological Survey, Sask. Miner. Resour., Misc. Rep. 78-10, p16-24

MacQuarrie, R.R. (1979): Geological re-investigations mapping, Birch Portage South (NTS area 63L-15S); *in* Summary of Investigations 1979, Saskatchewan Geological Survey, Sask. Miner. Resour., Misc. Rep. 79-10, p29-38.

Pyke, M.W. (1966): The geology of the Pelican Narrows and Birch Portage areas, Saskatchewan; Sask. Dep. Miner. Resour., Rep. 93, 68p.

Petrography, Fluid Inclusion, and Stable Isotope Data from the Bakos Lode Gold Deposit, La Ronge Domain[1]

M. Fayek[2], T.K. Kyser[2] and R.T. Kusmirski[3]

Fayek M., Kyser T.K. and Kusmirski, R.T. (1991): Petrography, fluid inclusion, and stable isotope data from the Bakos lode gold deposit, La Ronge Domain; *in* Summary of Investigations 1991, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 91-4.

The Bakos lode gold deposit is situated 45 km northeast of La Ronge, Saskatchewan, in the Proterozoic La Ronge Domain. The deposit occurs within the late northeast-trending, brittle-ductile Bakos Shear Zone. This shear zone cross-cuts the central granite and granodiorite phases of the composite Little Deer Lake Pluton (Figure 1) (Chapman *et al.*, 1990). The vein mineralogy, the size of the veins and the type and extent of alteration distinguishes this deposit from other epigenetic gold deposits in the La Ronge Domain.

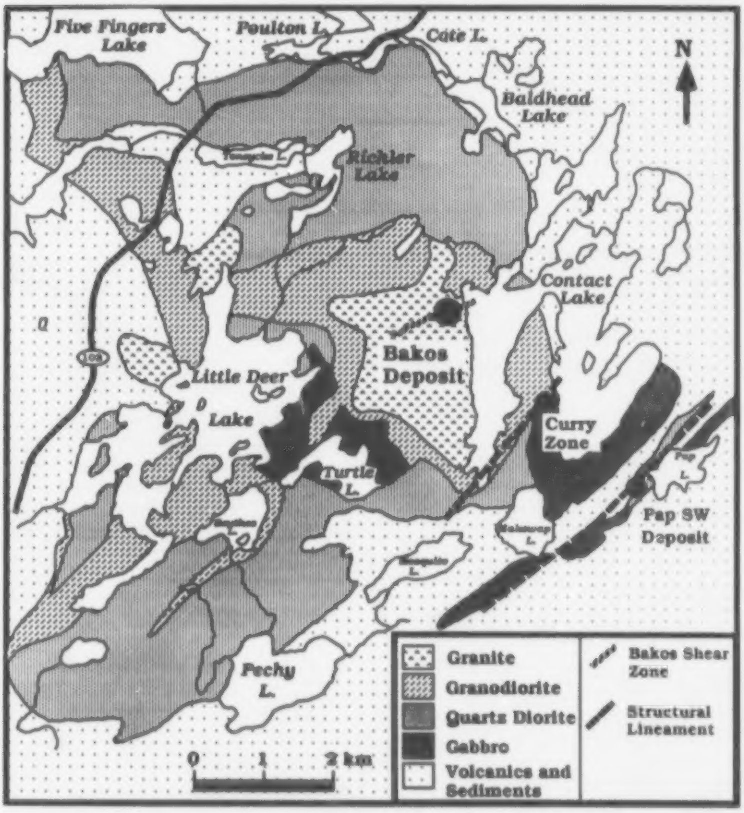


Figure 1 - Geology of the Little Deer Lake Pluton (modified from Chapman et al., 1990).

1. Petrography

In several recent studies of vein systems within brittle-ductile shear zones (i.e.: Duport Mine, Ontario (Smith, 1986), Sigma Mine, Quebec (Robert *et al.*, 1983), and various other deposits), veins occur in the D, P, R and R' shears. The interpretation of fabrics and vein geometry in terms of this shear zone model is important because it identifies the fundamental nature of the deformation process, and can aid in predicting the attitude and location of the ore shoots (Roberts, 1987).

According to this shear zone model, several vein types occur in the Bakos deposit (Figure 2). Both R and R' type veins crosscut the main Bakos shear foliation at low and high angles, respectively. These veins consist of quartz-muscovite ± microcline ± euhedral pyrite ± pyrrhotite. They are weakly to moderately deformed as indicated by quartz recrystallization along serrated grain boundaries of large deformed quartz grains and fractured euhedral pyrite. The R and R' veins are considered to be the first veins to have formed, followed by D vein formation in the central part of the shear zone along the C plane of foliation. D-type veins consist of quartz-muscovite ± pyrite ± pyrrhotite and are moderately to strongly deformed, exhibiting pervasive subgrain formation. Extension fractures crosscut D-type veins and consist mainly of quartz and fine-grained muscovite.

---

(1) This project is supported by a University-Industry grant from NSERC and Cameco.
(2) Department of Geological Sciences, University of Saskatchewan, Saskatoon, Saskatchewan, S7N 0W0
(3) Cameco Corporation, 2121 11th St. W., Saskatoon, Saskatchewan, S7M 1J3

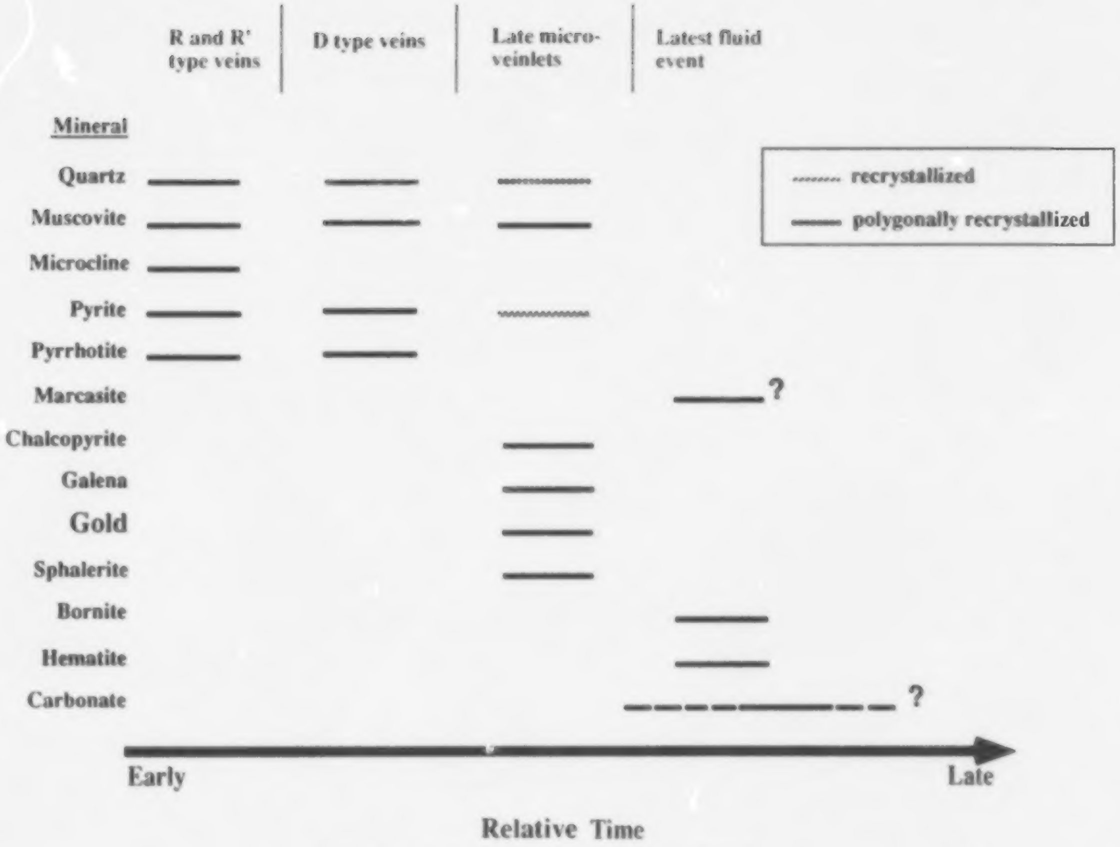


Figure 2 - Vein paragenesis from the Bakos deposit.

Microveinlets of recrystallized polygonal quartz with fine-grained muscovite ± recrystallized sieve-textured pyrite ± chalcopyrite ± galena ± sphalerite ± marcasite and gold, occur throughout D-, R-, and R'-type veins. The sulphides and gold occur as overgrowths and fracture fillings in the deformed, early euhedral pyrite. Gold also occurs as free gold associated with recrystallized polygonal quartz and fine-grained muscovite.

2. Fluid Inclusions

Fluid inclusion studies provide information about the composition, pressure, and temperature of the fluids present during mineralization and that may have subsequently interacted with the deposit. Measurements were made on several primary and secondary inclusions from recrystallized polygonal quartz associated with the gold. Primary inclusions occur as irregular clusters and isolated individual inclusions far removed from fractures or planar features. Secondary inclusions occur as broad planar arrays which crosscut grain boundaries.

Fluid inclusions are classified according to their appearance at room temperature. Preliminary results indicate that primary and secondary inclusions are similar and are H_2O-rich. Both inclusion types are aqueous, two phase inclusions with consistent liquid to vapour ratios. Homogenization temperatures, which represent minimum trapping temperatures, are between 220° and 300°C. Salinities of the fluid inclusions range from 1 to 6 weight percent NaCl equivalent, but typically average 4 weight percent NaCl equivalent.

3. Stable Isotopes

Quartz separates were collected from the Little Deer Lake Pluton, and from various types of veins from the Bakos deposit. The $\delta^{18}O$ values of quartz from the pluton are between 9.3 and 10.3 per mil and are $\delta^{18}O$-depleted relative to quartz from any of the veins (Figure 3). The $\delta^{18}O$ of vein quartz from the Bakos deposit has a relatively small range (11.4 to 13.0 per mil)(Figure 3); however, D-type veins, occurring in zones of intense deformation, are $\delta^{18}O$-rich relative to vein quartz from zones of weak to moderate deformation (R-type veins) and polygonal quartz.

Using quartz-muscovite fractionation factors (Table 1), the isotopic equilibrium temperatures were calculated from the $\delta^{18}O$ values of muscovite in textural equi-

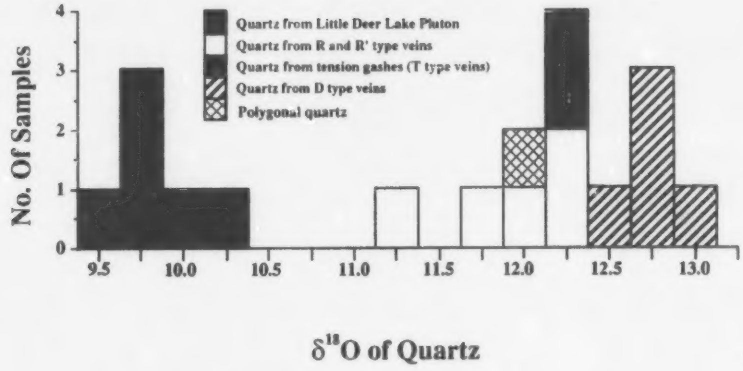


Figure 3 - Histogram of $\delta^{18}O$ values of quartz associated with the Bakos deposit.

Table 1: Calculated $\delta^{18}O$ Values and Temperatures for Fluids in Equilibrium with Minerals in the Bakos Deposit Veins

Sample No.	$\delta^{18}O$ Quartz	$\delta^{18}O$ Mineral	Temperature °C
119-13 D type qtz vein	12.8	musc. 9.2	295
119-40 D type qtz vein	12.7	musc. 9.4	310
91 R type qtz vein	12.4	musc. 9.2	330
69-C2 polyg. qtz	11.9	musc. 9.8	510

Fractionation factors used are cited in Kyser, (1987) and references there in.

Abbreviations: qtz = quartz, musc. = muscovite, polyg. = polygonal

Note: all stable isotope values are in per mil

librium with quartz, for D- and R-type veins, and microveinlets associated with the gold. Fluids from D- and R-type veins had a temperature of ~300°C whereas fluids from the microveinlets had much higher temperatures near 500°C.

4. Conclusions

The shear zone model described by Roberts (1987), has been tentatively applied to the Bakos Deposit in an attempt to classify various veins exhibiting a variety of orientations relative to the main Bakos trend. Alternating brittle and ductile behaviour is observed in the veins from the Bakos deposit. Enhanced ductility during shearing occurs at high temperatures because of high strain. Grain size reduction occurs which forms quartz subgrains and increases grain boundary flow. Consequently vein permeability increases allowing fluids to flow at grain boundaries. An increase in fluid pressure associated with this fluid can cause a decrease in the effective normal stress and thereby induce brittle failure at lower deviatoric stresses (Murphy, 1989). The hot (~500°C) recrystallizing fluid associated with the gold is envisioned to have permeated along zones of weakness created by brittle failure.

By combining fluid inclusion data and stable isotope geothermometry, primary and secondary fluid inclusions in recrystallized polygonal quartz associated with the gold give pressure estimates of 2.0 to 4.2 kbars. Similar studies from the Jasper Mine (Hrdy *et al.*, 1990) and the Star Lake Mine (Ibrahim and Kyser, in press) give pressure estimates of 3 to 5 kbars, and ore forming fluid temperatures as high as 500°C, which are similar to the auriferous fluids at the Bakos deposit.

5. References

Chapman, R., Curry, G. and Sopuck, V. (1990): The Bakos deposit discovery-A case history; *in* Beck, L.S. and Harper, C.T. (eds.), Modern Exploration Techniques, Sask. Geol. Soc., Spec. Publ. No. 10, p195-212.

Hrdy, F., Kyser. and Kusmirski, R.T. (1990): Fluid inclusion, stable isotope and microprobe data from the Jasper Zone lode gold deposit, La Ronge Domain; *in* Summary of Investigations 1990, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 90-4, p163-165.

Ibrahim, M.S. and Kyser, T.K. (in press): Fluid inclusion and isotope systematics of the high temperature Proterozoic Star Lake lode-gold deposit, northern Saskatchewan, Canada; Econ. Geol.

Kyser, T.K. (1987): Equilibrium fractionation factors for stable isotopes; *in* Kyser, T.K. (ed.), Short Course in Stable Isotope Geochemistry of Low temperature Fluids, Mineral. Assoc. Can., v13, p1-84.

Murphy, J.B. (1989): Tectonic environments and metamorphic characteristics of shear zones; *in* Mineralization and Shear Zones, Geol. Assoc. Can., short course notes, v6, p29-49.

Robert, F., Brown, A.C. and Audet, A.J. (1983): Structural control of gold mineralization at the Sigma Mine, Val d'Or, Quebec; Can. Instit. Min. Metall., Bull., v76, no.850, p72-80.

Roberts, R.G. (1987): Ore Deposit Models, Number 11: Archean lode gold deposits; Geosci. Can., v14, p37-52.

Smith, P.M. (1986): Duport, a structurally controlled gold deposit in northwestern Ontario, Canada; *in* MacDonald, A.J. (ed.), Proceedings of Gold '86, an International Symposium on the Geology of Gold, Toronto, 1986, p197-212.

Mineral Parageneses, Fluid Characteristics and Radiogenic Isotopic Data from the Proterozoic Jasper Gold Zone, La Ronge Domain[1]

F. Hrdy[2], T.K. Kyser[2] and R.T. Kusmirski[3]

Hrdy, F., Kyser, T.K. and Kusmirski, R.T. (1991): Mineral parageneses, fluid characteristics and radiogenic isotopic data from the Proterozoic Jasper gold zone, La Ronge Domain; *in* Summary of Investigations 1991, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 91-4.

A petrogenetic study of the Jasper gold zone was initiated in 1989 (Hrdy *et al.*, 1990) and is ongoing. Expansion and resolution of vein paragenesis and the application of fluid inclusion, stable and radiogenic isotope systematics has allowed derivation of fluid temperatures, pressures, sources, and possible ages. The past year of study has revealed the presence of a visible sericite alteration envelope which extends 2 to 3 m around the main structure. However, stable isotope data indicate that there is more laterally extensive cryptic alteration of the $\delta^{18}0$ value of quartz in the host pluton. Muscovite, which on the basis of petrographic relations, is paragenetically related to gold in its present form, yields an $^{40}Ar/^{39}Ar$ age of 1710 Ma. Single zircon $^{207}Pb/^{206}Pb$ dating yields an age of 1855 ±8 Ma for the hosting Island Lake intrusion as well as surprising complex and variable early Proterozoic to mid-Archean ages for the altered margins of zircons hosted within the auriferous veins.

1. Vein Paragenesis

Petrography and electron microprobe examination of over 25 polished thin sections has revealed a complex mineral paragenesis and extensive mineral inventory from auriferous quartz veins (Figure 1). The concentration of minerals in the veins varies locally, but averages approximately 70 to 80 percent quartz and 5 to 20 percent pyrite with minor muscovite, biotite, chlorite, tourmaline, albite, carbonate, sphalerite, chalcopyrite, galena, marcasite, and trace gold, silver-gold tellurides, molybdenite, pyrrhotite, covellite, tetrahedrite, scheelite, monazite, rutile, and zircon. Uranium-rich solid inclusions in altered zircon rims have also been observed.

Quartz deposition has a prolonged history coincident with, and outlasting, structural deformation, as indicated by quartz grain recrystallization and cross-cutting relations. Gold occurs within cross-cutting fractures and microfractures which transect all other features, and is associated with recrystallized quartz subgrains. This indicates a relatively late introduction (or remobilization) of gold where pressure reduction, either within a brittle fracture or due to grain size reduction (volume decrease), caused gold precipitation. Brittle microfractures in rafted wallrock clasts in the veins host silver-gold tellurides (Figure 2) and are sites where potassium feldspar alters to albite. Silver-gold tellurides may be a manifestation of interaction between the auriferous fluid and the surrounding lithology and not necessarily a separate depositional event. Minerals paragenetically related to gold deposition include: muscovite, sphalerite, galena, minor pyrite, and trace monazite, molybdenite, scheelite, albite, and chlorite (Figure 1).

2. Fluid History

A formation temperature of 335°C was derived from quartz and muscovite oxygen isotope systematics (Eslinger, *et al.*, 1979). Muscovite is paragenetically associated with gold deposition and the quartz was selected on the basis of enriched gold content and development of grain size reduction. Due to complex mineral and fluid inclusion paragenesis, a pressure history for the Jasper vein is still unknown. Calculated $\delta^{18}0$ and δD values for fluids responsible for vein formation indicate a metamorphic source. Fluid inclusions associated with gold at recrystallized quartz subgrain boundaries and within healed fractures are generally two-phase with a large vapour bubble and typically a liquid CO_2 outer rim. Methane and other gas components are indicated and high, but variable, bulk homogenization/decrepitation temperatures (290° to 580°C) were obtained. Heterogeneous trapping, as a result of fluid-CO_2 separation, is suggested to explain the variation of fluid inclusion compositions. This would also provide a mechanism by which gold deposition could have occurred.

3. Alteration

The extent of wallrock-fluid interaction which may have occurred at Jasper has not been documented. Alteration of wallrock mineralogy is petrographically visible only within 2 to 3 m of the shear zone and consists of minor sericite and carbonate filling fractures and as overgrowths. The $\delta^{18}0$ values of quartz obtained from drill holes which transect representative portions of the wallrock and mineralization indicate an alteration envelope of greater than 10 m laterally from the main zone. The $\delta^{18}0$ values of the quartz from the host rock is up to 2 per mil greater than quartz in unaltered por-

---

(1) This project supported by university-industry grant from NSERC and Cameco.

(2) Department of Geological Sciences, University of Saskatchewan, Saskatoon, Saskatchewan, S7N 0W0.
(3) Cameco Corporation, 2121 11th St. W., Saskatoon, Saskatchewan, S7M 1J3.

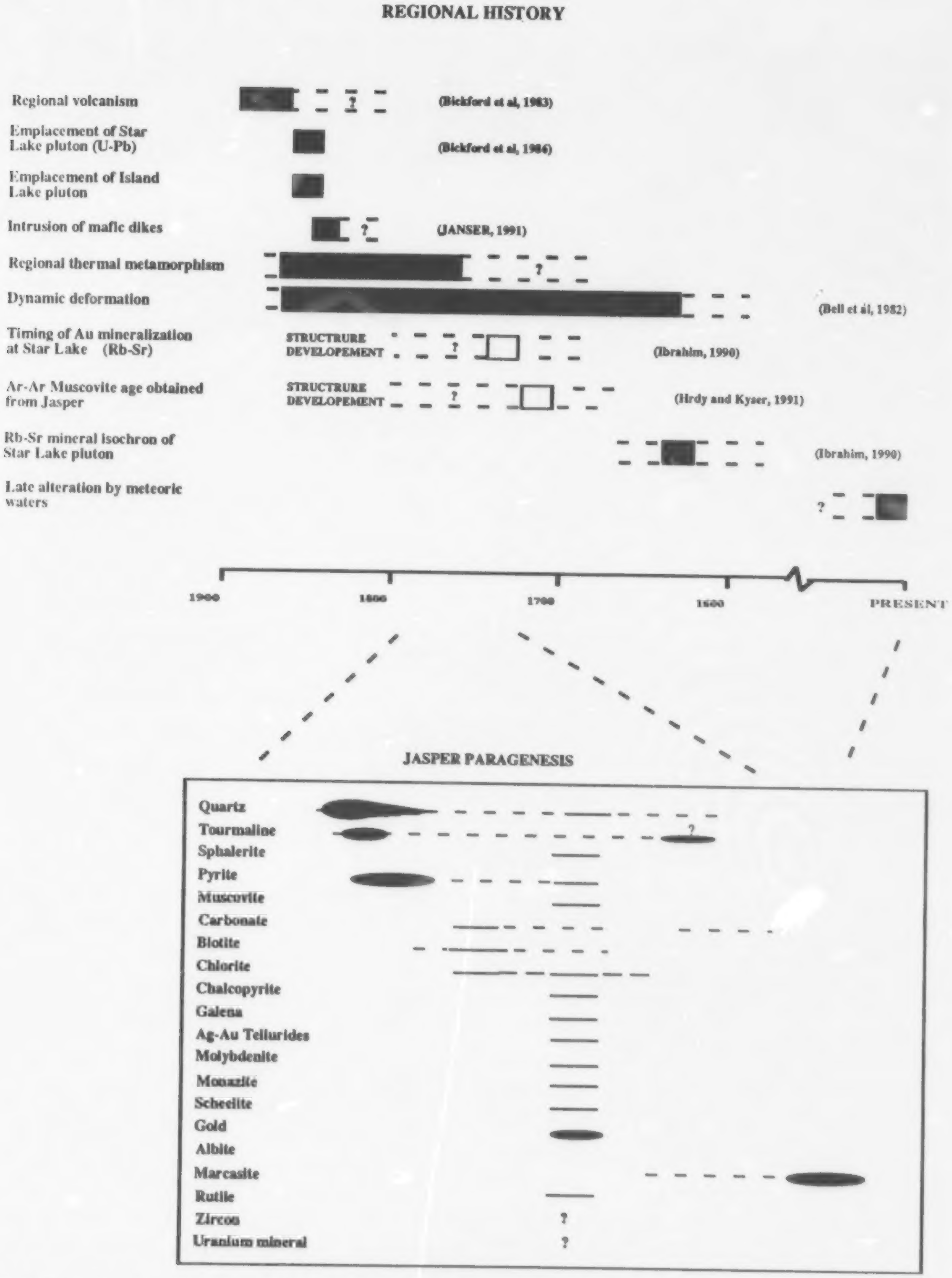

REGIONAL HISTORY
Regional volcanism
?
(Bickford et al, 1983)
Emplacement of Star Lake pluton (U-Pb)
(Bickford et al, 1986)
Emplacement of Island Lake pluton
Intrusion of mafic dikes
?
(JANSER, 1991)
Regional thermal metamorphism
?
Dynamic deformation
(Bell et al, 1982)
Timing of Au mineralization at Star Lake (Rb-Sr)
STRUCTRURE DEVELOPEMENT
?
(Ibrahim, 1990)
Ar-Ar Muscovite age obtained from Jasper
STRUCTRURE DEVELOPEMENT
?
(Hrdy and Kyser, 1991)
Rb-Sr mineral isochron of Star Lake pluton
(Ibrahim, 1990)
Late alteration by meteoric waters
?
1900
1800
1700
1600
PRESENT
JASPER PARAGENESIS
Quartz
Tourmaline
?
Sphalerite
Pyrite
Muscovite
Carbonate
Biotite
Chlorite
Chalcopyrite
Galena
Ag-Au Tellurides
Molybdenite
Monazite
Scheelite
Gold
Albite
Marcasite
Rutile
Zircon
?
Uranium mineral
?

Figure 1 - Summary of geologic events for the La Ronge Domain and relative ages for vein mineralogy from the Jasper auriferous structure.

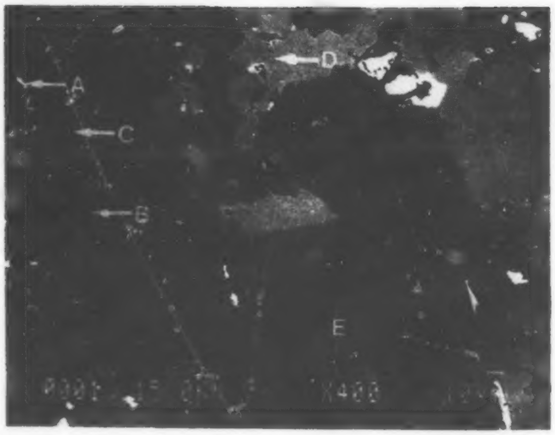


Figure 2 - S.E.M. photograph of microfractures within rafted wallrock in the vein hosting silver-gold tellurides. Muscovite and chlorite are also present. Darker regions coincident with fractures indicate potassium feldspar alteration to albite. A = silver-gold telluride; B = chlorite in microfracture; C = K-feldspar altered to albite; D = K-feldspar; E = tourmaline.

tions of the pluton. Quartz in the veins has $\delta^{18}O$ values of 12 to 14 per mil whereas quartz from the pluton has much lower values of 8 to 9 per mil. As shown in Figure 1, the bulk of the quartz in the veins is not associated with the fluid that precipitated the gold so that the elevated $\delta^{18}O$ values resulting from alteration do not necessarily reflect the presence of an auriferous fluid.

4. Radiogenic Studies

$^{39}Ar/^{40}Ar$ dating of vein-derived muscovite yielded a date of 1710 Ma. The single zircon $^{207}Pb/^{206}Pb$ evaporation technique, as described by Kober (1986, 1987) and Ansdell and Kyser (1990), provides an age of 1855 ±8 Ma for the Island Lake intrusion. Zircons have also been recovered from mineralized veins. During step heating of vein zircons, $^{207}Pb/^{206}Pb$ ages as old as 3 Ga become progressively younger as temperatures are increased. This indicates the presence of older rims on the zircons which are a result of a reaction of the zircons with hydrothermal fluids that have interacted with Archean basement.

5. Conclusion

The mineral inventory of the auriferous veins at Jasper is extensive and their paragenesis has been determined. Gold deposition occurred during vein deformation, but relatively late in the overall history of the structure. Fairly extensive cryptic alteration of host rock is suggested and may indicate a high fluid/wallrock ratio during deposition of the majority of the quartz. Paragenetically, quartz deposition occurred prior to the introduction of auriferous fluids. Radiogenic dating supports late introduction of gold and provides evidence that Archean basement may have been one of the sources of the late auriferous fluids.

6. References

Ansdell, K.M. and Kyser, T.K. (1990): Age of Granitoids from the Flin Flon Domain: An Application of the Single-zircon Pb-evaporation Technique; *in* Summary of Investigations 1990, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 90-4, p136-142.

Bell, K. and Macdonald, R. (1982): Geochronological calibration of the Precambrian Shield in Saskatchewan; *in* Summary of Investigations 1982, Saskatchewan Geological Survey, Sask. Miner. Resour., Misc. Rep. 82-4, p17-22.

Bickford, M.E., Lewry, J.F. and Van Schmus, W.R. (1983): Zircon isotope geochronology project; U-Pb (zircon) chronology for the Hudsonian orogenic terrains in Saskatchewan; *in* Summary on Investigations 1983, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 83-4, p34-39.

Bickford, M.E., Van Schmus, W.R., Macdonald, R., Lewry, J.F. and Pearson, J.G. (1986): U-Pb zircon geochronology project for the Trans-Hudson orogen: current sampling and recent results; *in* Summary of Investigations 1985, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 86-4, p101-107.

Eslinger, E.V., Savin, S.M. and Yeh, H. (1979): Oxygen isotope geothermometry of diagenetically altered shales; SEPM Spec. Publ. No. 26, p113-124.

Hrdy, F., Kyser, T.K. and Kusmirski, R.T. (1990): Fluid inclusion, stable isotope and microprobe data from the Jasper Zone lode gold deposit, La Ronge Domain; *in* Summary of Investigations 1990, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 90-4, p163-165.

Ibrahim, M.S. (1990): History of the fluids associated with the Proterozoic Star Lake loge-gold deposit, northern Saskatchewan, Canada; unpubl. MSc. thesis, Univ. Sask., 124p.

Kober, B. (1986): Whole-grain evaporation for $^{207}Pb/^{206}Pb$-age-investigations on single zircons using a double-filament thermal ion source; Contrib. Mineral. Petrol., v93, p482-490.

__________ (1987): Single-grain evaporation combined with Pb+ emitter bedding for $^{207}Pb/^{206}Pb$-age-investigations using thermal ion mass spectrometry, and implications to zirconology; Contrib. Mineral. Petrol., v96, p63-71.

Tourmaline, Phosphate Minerals, Zircon and Pitchblende in the Athabasca Group: Maw Zone and McArthur River Areas, Saskatchewan[1]

D. Quirt[2], T. Kotzer[3] and T.K. Kyser[3]

Quirt, D.H., Kotzer, T. and Kyser, T.K. (1991): Tourmaline, phosphate minerals, zircon, and pitchblende in the Athabasca Group: Maw Zone and McArthur River areas, Saskatchewan; *in* Summary of Investigations 1991, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 91-4.

Anomalously high abundances of several incompatible-element-rich minerals have been found in the Helikian Athabasca Group sandstone and conglomerate in the McArthur River-Wheeler River area (Assessment Files 74H-0006, -0010, -0011). Dravite, hematite, phosphate minerals, clay minerals and euhedral quartz occur in alteration halos surrounding Athabasca unconformity-type U±Ni±Co±As±Cu±S mineralization in this area (Hoeve and Quirt, 1984; Kotzer and Kyser, 1990a). A similar alteration assemblage of xenotime, dravite and euhedral quartz occurs in altered sandstones hosting heavy rare earth element (HREE), yttrium-rich phosphate mineralization at the Maw Zone, near Brook Lake (Agip Canada Ltd., 1985; MacDougall, 1990).

The objectives of the present investigation are:

1) to illustrate and characterize the variety of rare earth element (REE)-bearing minerals found in the basal portion of the Athabasca Group;

2) to document the geochemical composition of the dravite; and

3) to determine whether the quartz-dravite alteration assemblages evident in both the unconformity-type uranium mineralization and the HREE-yttrium phosphate mineralization have formed from similar fluids and processes.

To determine 3), the chemical and REE compositions of diagenetic phosphates and tourmalines in the altered sandstones hosting the HREE-yttrium phosphate mineralization are compared to the chemical and REE compositions of phosphates, tourmalines and uraninite from areas of high grade uranium mineralization at McArthur River (P2 Main, BJ, P2 North) where the fluid history and mineral paragenesis is known. Results of these comparisons will then be used to; 1) ascertain which minerals are hosting the rare-earth elements and yttrium, 2) constrain the origin and timing of fluids associated with the HREE-yttrium phosphate mineralization relative to the fluids related to the unconformity-type uranium deposits and, 3) suggest sources of the REE and mechanisms for the transport and fractionation of REE into the various minerals.

Throughout the course of this study, diagenetic and detrital tourmalines and diagenetic phosphate minerals in samples of silicified and tourmalinized Manitou Falls Formation sandstones from the Maw Zone and the McArthur River area were examined by transmitted light microscopy. Their major element compositions were determined with an electron microprobe; REE compositions in dravite, pitchblende, phosphate and clay minerals were determined by electron microprobe and ICP-MS.

1. Geological Setting and Alteration

The Athabasca Group is composed primarily of first cycle fluviatile and marine quartz sandstone and conglomerate. In the southeastern portion of the Athabasca Basin (the location of the study area) the basal sequence of the stratigraphic column consists of the fluviatile Manitou Falls Formation. All sandstones hosting the tourmaline and phosphate minerals analyzed in this study are from this formation. It is composed of permeable, medium- to coarse-grained conglomerate and pebbly sandstone, overlain by sandy braided stream and wadi deposits. Mineral grains in these rocks are subrounded to well rounded detrital quartz.

Diagenetic minerals present in the sandstone include kaolinite, illite, hematite, goyazite, tourmaline and quartz (Hoeve and Quirt, 1984). Hematite is ubiquitous giving purple to reddish to pink colouration to the rock. Superimposed diagenetic chemical reduction has resulted in zones having a bleached or mottled appearance. Goyazite, a Ca-Sr-LREE-phosphate, is found in trace quantities throughout the Athabasca Group. Diagenetic Mg-tourmaline (dravite) is found in pore spaces as wispy tufts of radiating needles originating from surface defects on detrital quartz grains and as oriented overgrowths on detrital tourmaline grains. Detrital heavy mineral bands are common and are characterized by hematite, rutile, zircon, and minor tourmaline and monazite.

---

(1) Project funded in part by Saskatchewan Research Council (D.H.Q.) and NSERC Operating grant to T.K.K.
(2) Saskatchewan Research Council, 15 Innovation Blvd., Saskatoon, Saskatchewan, S7N 2X8.
(3) Department of Geological Sciences, University of Saskatchewan, Saskatoon, Saskatchewan, S7N 0W0.

Maw Zone samples consist of pinkish and variably-bleached hematitic sandstones, sandstone-euhedral quartz breccias, and silicified emerald green sandstone. Petrographically, alteration of the sandstones consists of quartz overgrowths, radially-textured clear to green aggregates of interstitial acicular dravite, euhedral quartz, altered brownish-green detrital tourmaline, granular xenotime, phosphate-rich altered detrital zircon, and fractured, detrital zircons rimmed by xenotime. Negligible amounts of illite and kaolinite are present. In the highly brecciated silicified sandstones, the matrix consists mainly of euhedral quartz and radially textured dravite surrounding altered sandstone blocks.

The Athabasca Group in this area is underlain by Wollaston Domain, Archean granitoid gneisses and Aphebian Wollaston Group metasediments. The northeast-trending metasediments comprise an interbedded sequence of meta-quartzite, feldspathic meta-quartzite, and pelitic to psammopelitic to psammitic gneiss, with minor amphibolite and pegmatite. A sub-Athabasca paleoridge associated with meta-quartzite is spatially associated with the uranium mineralization at McArthur River and the REE mineralization at the Maw Zone. The basement rocks have undergone several episodes of folding followed by several episodes of ductile to brittle thrusting and faulting, commonly localized along contacts of graphitic gneiss with the Archean granitoid. The faults have remained active to recent times (Wilson and Kyser, 1987; Annesley, 1989; Annesley and Madore, 1988, 1989; Kotzer and Kyser, 1991). Peak metamorphic conditions are interpreted as being middle amphibolite to lower granulite grade.

a) McArthur River

Highly altered hematitic and bleached sandstones and euhedral quartz sandstone breccias near uranium mineralization from the McArthur River area (P2 main, P2 North, BJ) contain significant amounts of euhedral quartz and tourmaline. They are petrographically similar to those of the Maw Zone, consisting of an alteration assemblage of quartz overgrowths, varying amounts of euhedral quartz and dravite, goyazite intergrown with dravite and illite in pore spaces, fractured detrital zircons, hematite and rutile.

At the unconformity between the Athabasca Group and the sub-Athabasca lithologies is a well developed paleoweathering profile. This profile consists of kaolinite with hematite at the top grading downwards, through less altered rocks containing kaolinite-illite, to mildly altered rocks with a chlorite-illite assemblage to fresh rock.

Several episodes of mixing of fluids from the basement with fluids in the sandstones have produced and modified the unconformity-type uranium mineralization (Hoeve and Sibbald, 1978; Hoeve and Quirt, 1984; Sibbald, 1985; Wilson and Kyser, 1987; Kotzer and Kyser, 1990a). Associated with these fluid mixing events was neoformation of minerals such as illite, Al-Mg chlorite, dravite, euhedral quartz, and hematite, which overprint the diagenetic mineral assemblages. A later retrograde assemblage of kaolinite with chemically and isotopically altered illite and chlorite overprints the diagenetic and prograde mineral assemblages in the vicinity of fault and fracture zones (Wilson and Kyser, 1987; Kotzer and Kyser, 1991).

b) Maw Zone

The Maw Zone is expressed on surface by a silicified, dravitized, brecciated, faulted, and anomalously radioactive sandstone outcrop. Outcrop geology has been described and mapped by Agip Canada Ltd. (1985) and MacDougall (1990). REE mineralization was discovered in 1982 by Agip Canada Ltd., who subsequently obtained yttrium values greater than 2.0 percent, boron values up to 2.5 percent, and total rare earth oxide (REO) values up to 8.1 percent.

MacDougall (1990) interpreted the brecciated sandstone to be a result of clastic depositional processes that fragmented the sediments soon after deposition or somewhat later when more consolidated. The brecciated zone is not only present at surface but can be traced downwards through the sandstone below the outcrop.

Detailed logging of core from Agip drill hole ZQ-10-83 (Quirt, unpublished data; Agip Canada Ltd., 1985) indicates that zones of strong brecciation and sandstone blocks, up to 5 m in thickness, with rotated bedding extend from surface to nearly 200 m downhole. The breccia zones are commonly healed by dravite and illite. Below this depth, to within about 10 m of the unconformity (at 256.3 m), bedding is locally highly disturbed, folded and faulted. In addition to the brecciation, fracturing of the sandstone is ubiquitous, both on the outcrop as noted by MacDougall (1990), and in drill core. The upper 170 m of this drill core is pervasively silicified and fractured, the fractures commonly being open and lined with euhedral quartz. Those fractures not lined with euhedral quartz are healed with silica and dravite. The large vertical extent and restricted horizontal extent (Agip Canada Ltd., 1985) of the brecciation and fracturing at the Maw Zone indicates that the process responsible is more likely one of intense hydrothermal activity, as has been described for the McArthur River uranium deposit (Hubregtse and Sopuck, 1987; Kotzer and Kyser, 1990a, 1991), than one of syn- to early post-depositional fragmentation.

The Maw Zone is on the east side of a northeast-trending reverse fault, that dips to the southeast. The fault is believed to occur at the contact between Archean granitic gneiss (to the northwest) and Wollaston Group pelitic to psammopelitic graphitic gneiss. The fault is on the north(west) side of the basement topographic high, formed by meta-quartzite and psammopelitic to pelitic gneiss, that underlies the Maw Zone (Agip Canada, 1985). This sub-Athabasca ridge was interpreted to be an eroded pre-Athabasca thrust block. Diamond drilling indicates that the ridge is 150 m or more above its surroundings and extends for approximately 8.5 km, the presumed extent of the meta-quartzite unit and the fault. The peak of the ridge is about 0.5 km in length, in the region of the Maw Zone.

Post-Athabasca thrusting was not considered by Agip Canada Ltd. (1985), therefore, no note was made of the possibility of the presence of thrust slices or interleaving of basement lithologies with the Athabasca sandstone. Preliminary petrographic, mineralogical, and geochemical examinations of two clay mineral-rich and hematite-rich intervals in drill hole ZQ-10-83 indicate that regolithic Wollaston Group gneiss may be present as thrust slices in the Athabasca sandstone below the Maw Zone.

2. Petrography and Chemistry

a) Diagenetic dravite

Unique to the Maw Zone is the presence of 'emerald' dravite which imparts a light green to olive green to emerald green colour to the sandstone and siltstone. MacDougall (1990) noted the presence of olive green quartz grains, however, this coloration is commonly due to the presence of 'emerald' dravite as a matrix mineral. In outcrop, only patchy zones of greenish sandstone are found (cf map in MacDougall, 1990). In drill core, greenish to bluish-green dravite is most prevalent as a fracture filling and as breccia matrix within pinkish to pinkish-buff to bleached-white sandstone in the upper 125 m of the core (i.e. more than 130 m above the unconformity). Within this interval, extensive development of pale green to emerald-green pervasive dravitization is restricted to relatively thin zones of less than 5 m in width. Below this interval, variably pervasive dravitization of the sandstone, plus dravitic fracture fillings and breccia matrix, imparts a pale blue to bluish-grey colour to the rock.

Diagenetic tourmaline in the euhedral quartz-breccias and altered hanging-wall sandstones from the McArthur River area (Plate 1- typical of P2 main, P2 north, BJ) and from the euhedral quartz-sandstone breccias and emerald-green sandstones from the Maw zone (Plates 2 and 3) have Fe/Mg ratios and Al_2O_3 contents (#1 in Table 1) corresponding to an alkali-free dravite (Deer *et al.*, 1982). 'Emerald' dravite in the silicified emerald-green sandstone from the Maw Zone, however, is significantly enriched in Cr_2O_3 (#2 - Table 1). Based on chromium and magnesium content, it can be classified as chrome dravite (Dunn, 1977).

Table 1 - Chemical Compositions of Tourmaline from the Maw Zone and the McArthur River Area.

group	#1	#2	#3	#4	#5
n	4	1	3	2	1
SiO_2 (%)	37.89	38.54	38.07	37.74	37.96
Al_2O_3 (%)	37.44	36.73	31.82	37.46	37.71
TiO_2 (%)	0.01	0.01	0.89	<0.01	0.02
FeO (%)	0.70	0.74	6.18	1.18	0.79
MnO (%)	0.02	0.02	0.02	<0.01	0.04
MgO (%)	8.58	8.31	8.23	8.28	7.95
Cr_2O_3 (%)	0.03	1.68	0.01	<0.01	1.19
CaO (%)	0.17	0.13	1.24	0.15	0.22
Na_2O (%)	0.85	0.80	1.94	0.70	0.51
K_2O (%)	0.02	0.01	<0.01	<0.01	0.01
NiO (%)	0.01	0.01	0.05	0.01	<0.01
Fe/Mg	0.11	0.12	1.01	0.19	0.13

Chemical compositions were determined for detrital tourmalines that display various degrees of alteration. Petrographically and chemically unaltered areas on the greenish-brown detrital tourmalines have Fe/Mg ratios and Al_2O_3 contents corresponding to a schorl-dravite (#3 - Table 1; Deer *et al.*, 1982). No significant amounts of Cr_2O_3 were detected. Altered detrital schorl-dravites, which appear as darker coloured areas (Plate 4), are substantially different with Fe, Mg and Al concentrations (#4 - Table 1) similar to the diagenetic alkali-free dravites in the pore spaces. Altered areas on the detrital schorl-dravites from the silicified emerald-green sandstone have similar Fe, Mg and Al contents as the altered schorl-dravites found elsewhere; however, they also contain elevated amounts of chromium (#5 - Table 1).

b) REE Mineralization

The REE mineralization at the Maw zone is microscopic in nature. It is not directly associated with the open fracturing or highly dravitized zones, but is contained within the zone of brecciated sandstone (Agip Canada Ltd., 1985). All samples containing REE mineralization are anomalously dravitic. Partial whole-rock geochemical analyses are presented in Table 2. In thin section the mineralization is indicated by the presence of xenotime, a Y-HREE-phosphate, as microcrystalline-granular, yellowish red-brown grains surrounded by acicular dravite, interstitial to quartz. Dravite appears to be intimately intermixed with the xenotime (Plate 2). In samples containing xenotime, Y, P, and REE are also found in significant quantities in altered detrital zircon.

Partial microprobe analyses of xenotime, zircon, and detrital monazite are presented in Tables 3 and 4. It is assumed that the difference between 100 percent and the oxide total is due to water and that other light elements that may be present in trace quantities (Deer *et al.*, 1962; Smith *et al.*, 1991).

The Y_2O_3 content of Maw Zone xenotime is approximately 40 weight percent, somewhat lower than for Alpine xenotime (Table 3). Water contents appear to be high (4.5 to 10.1 weight percent) as compared to the nearly stoichiometric Alpine xenotime (weight percent). The atomic ratio 100Y/(Y+REE) is almost constant at about 78 percent, falling within the upper range of reported values (Table 3; Jefford, 1962; Fieischer, 1989; Demartin *et al.*, 1991). Individual REO contents are almost constant and total around 19 weight percent (Table 3). Maw Zone xenotime is notably enriched in HREE, especially Dy, Gd, Er, and Yb, and, in addition, the distribution of the individual REO is distinctly different from those reported for fracture filling in the Alps (Table 4).

c) Altered zircon

Microprobe analyses of altered detrital zircon in samples that contain discrete xenotime indicates that significant quantities of Y, P, H_2O, and to a lesser extent U, Al, and

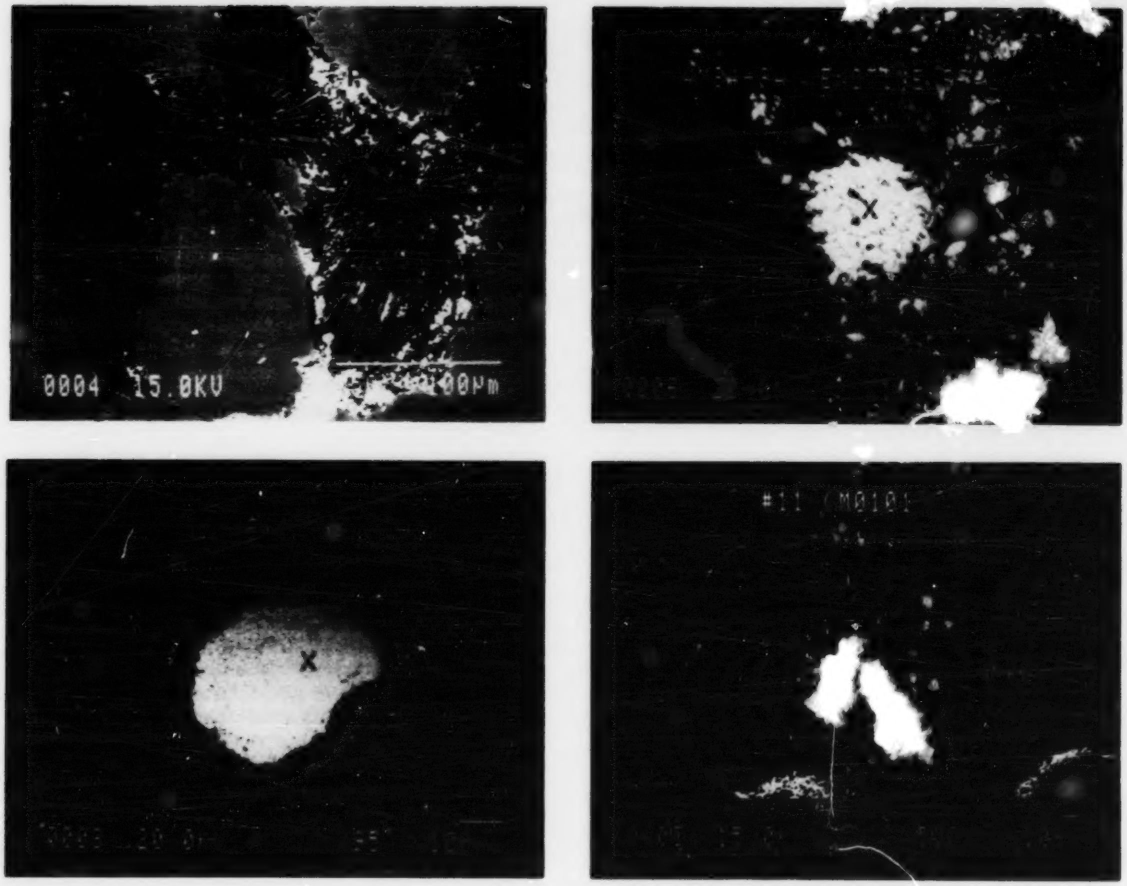


Plate 1 (upper left) - Back-scattered electron image of rosettes of acicular diagenetic tourmaline (dravite) in sandstone, McArthur River area. Q, quartz; P, goyazite; T, alkali-free dravite. Plates 2 (upper right) and 3 (lower left) - Back-scattered electron images of dravite and xenotime in sandstone, Maw Zone. Q, quartz; X, xenotime; T, dravite. Plate 4 (lower right) - Back-scattered electron image of altered detrital tourmaline in sandstone, Maw Zone. Q, euhedral quartz; Tsd, detrital schorl-dravite; Tc, chrome dravite.

Ca, are present (Table 3). The calculated amount of H_2O reaches 13 weight percent and the contribution of elements other than Zr, Si, and Hf range from 7 to 70 weight percent. Unaltered zircons from the Maw Zone area contain, near-stoichiometric proportions of Zr and Si, and low quantities (consistently less than 1.1 percent and commonly near zero) of the additional elements. Altered zircons vary from being indistinguishable from unaltered zircon (Plate 5) to being turbid and obviously altered (Plate 6), much like the granular xenotime.

The compositional information obtained from Maw Zone xenotime and zircon indicate the degree to which the zircon structure can incorporate foreign ions. In the altered zircons, Zr is mainly replaced by Y, REE, U, Ca, whereas Si is mainly replaced by P, H_2O, and, to some extent, Al. Zircon commonly contains Y due to the isostructural relationship of xenotime and zircon (Palache *et al.*, 1951; Deer *et al.*, 1962, 1982). Appreciable quantities of water may also be accommodated by the zircon structure (cyrtolite; Deer *et al.*, 1962, 1966, 1982). A general correlation appears to be present between water content and Y+P+REE content (Table 3; Deer *et al.*, 1962, 1982).

Unaltered detrital zircon and monazite (Tables 3, 4) are found within and surrounded completely by unfractured detrital quartz grains. Altered detrital zircon is found within fractured quartz grains (Plates 5 and 7) or as discrete detrital grains (Plate 8). The progression from unaltered zircon, with near-stoichiometric proportions of Zr, Si, and Hf, through increasingly altered zircon, with significant proportions of Y, P, and HREE, to xenotime is displayed in samples from the Maw Zone. The zircon analysis that returned sub-equal proportions of Zr, P, and Y, with some Si (Table 3, M-0101c), may represent an intergrowth of host zircon and replacement xenotime.

Table 2 - Selected Chemical and Mineralogical Analyses: Maw Zone and Wheeler River Area. Oxides in percent, elements in ppm; [1] RA- and ZQ-series data from D. Quirt, S.R.C. unpublished data file, (1984); Agip Canada Ltd. (1985); [2] samples courtesy of Cameco Corp.

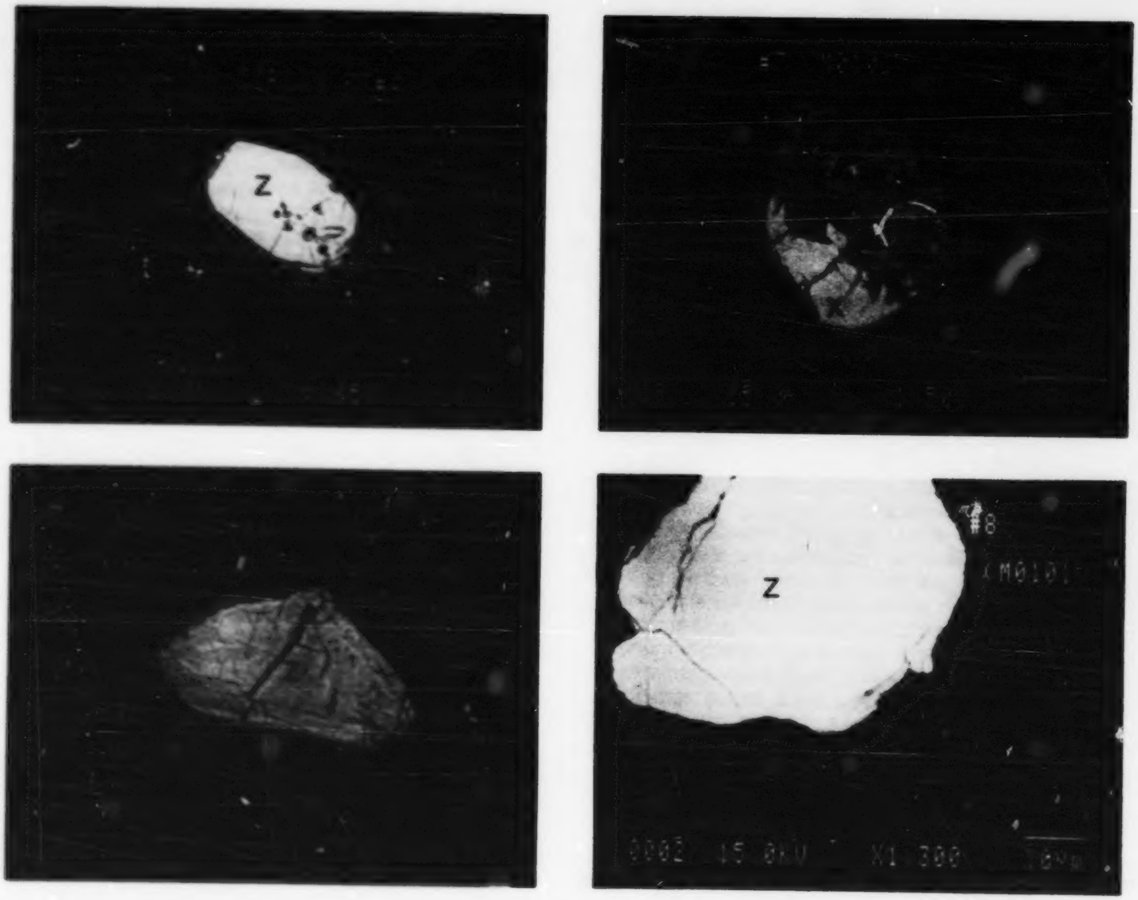
sample #	description	I+C+K=100			clay min.	dravite (0 to 6)	xenotime	Fe_2O_3(t)	P_2O_5	TiO_2	U	Ni	As	B	Sr	Zr	V	Au (ppb)	Y	La	Ce	Nd	Yb	sum REE	Yb/Ce ratio
		Illite	Chl.	Kaol.																					
AM-84-01	silicified sst.	0	0	100	yes	2	no	0.03	0.02	0.03	0.4	3	0.2	170	28	69	7	16	1	10	18	4	0.4	38	0.02
AM-84-02	clay pebble	0	0	100	yes	4	no	0.03	0.02	0.01	1.0	6	0.2	840	24	33	23	9	7	8	11	1	0.6	26	0.05
AM-84-03	'emerald' sst.	100	0	0	yes	2	no	0.02	0.02	0.02	0.8	4	0.3	420	28	34	14	4	3	10	13	4	0.3	31	0.02
AM-84-04	silicified sst.				no	2	no	0.09	0.02	0.02	2.2	251	33.0	260	22	58	10	38	2	21	17	5	0.4	47	0.03
AM-84-05	silicified sst.	0	36	64	yes	1	no	0.02	0.02	0.02	1.6	5	0.5	30	21	33	5	6	4	13	12	5	0.3	36	0.03
AM-84-06	'emerald' sst.				no	3	no	0.04	0.02	0.03	1.2	4	0.3	880	31	114	24	4	3	5	18	4	0.6	32	0.04
AM-84-09	dravitic sst.	0	0	100	yes	5	yes	0.10	0.07	0.02	5.4	6	0.8	3500	27	143	36	4	381	7	20	7	15.0	138	0.75
AM-84-10	dravitic sst.				no	3	no	0.04	0.02	0.03	1.0	3	0.8	700	38	82	10	5	9	13	24	6	0.8	50	0.03
AM-84-11	silicified sst.	0	45	55	yes	1	no	0.02	0.02	0.02	1.6	3	0.3	60	16	29	6	3	4	11	11	4	0.3	31	0.03
AM-84-12	'emerald' sst.				no	3	no	0.03	0.02	0.04	0.4	7	0.3	490	32	35	14	6	7	11	19	8	0.6	44	0.03
AM-84-13	regolith?	7	0	93	yes	1	no	3.02	0.29	1.09	8.6	11	3.7	100	962	2216	38	35	25	162	396	153	4.9	738	0.01
AM-84-14	regolith?	68	0	32	yes	5	no	0.91	0.19	0.63	6.6	23	1.6	12000	470	553	18	7	38	160	363	176	5.5	734	0.02
AM-84-15	'emerald' breccia				no	3	no	0.05	0.02	0.01	1.8	2	0.3	590	25	30	8	4	3	8	17	5	0.4	36	0.02
RA-1-83 [1]	silicified sst.	0	0	100	yes	4	yes	0.57	0.30	0.06	220.0	27	16.2	1570	39.1	442	72		36400	60.0	80	55	1100.0	7896	13.75
RA-2-83	clay-altered sst.						yes	0.49	0.12	0.03	420.0	36	12.5	599	17.2	208	62		19900	10.0	40	19	600.0	4763	15.00
RA-3-83	silicified sst.						yes	0.80	0.13	0.03	9.1	40	5.5	260	14.2	173	67		4000	6.7	14	8	110.0	957	7.86
RA-4-83	silicified sst.						yes	0.99	<0.01	0.02	16.4	55	1.8	213	17.1	33	78		500	8.1	19	9	1.7	52	0.09
RA-5-83	clay-altered sst.						yes	0.45	0.49	0.08	41.2	24	15.0	696	19.6	619	62		14600	9.0	32	8	400.0	3849	12.50
RA-6-83	sst. breccia						yes	0.46	0.24	0.04	70.4	25	20.0	400	10.8	311	67		14300	10.0	39	18	370.0	3527	9.49
RA-7-83	silicified sst.	4	0	96	yes	5	yes	0.47	0.51	0.07	160.0	19	23.8	3410	22.6	579	123		67600	70.0	100	35	1700.0	14245	17.00
ZQ-10-83-CA-02	silicified sst.	0	0	100	yes	2	no	0.53	0.02		6.0	9	2.3	797					105						
ZQ-10-83-CA-04	silicified breccia	0	0	100	yes	2	no	0.54	0.02		0.4	8	0.8	645					12						
ZQ-10-83-CA-06	silicified sst.	53	0	47	yes	2	no	0.66	0.02		0.7	17	1.5	519					14						
ZQ-10-83-CA-08	silicified sst.	99	0	1	yes	1	no	0.61	0.02		0.5	12	1.0	329					<1						
ZQ-10-83-CA-10	silicified sst.				no	3	no	0.62	<0.01		0.3	13	1.5	2470					3						
ZQ-10-83-CA-12	silicified sst.				no	2	no	0.50	<0.01		<0.2	11	2.1	924					<1						
ZQ-10-83-CA-14	silicified sst.	100	0	0	yes	4	yes	0.58	0.05		0.6	10	4.0	5430					445						
ZQ-10-83-CA-16	sst.	99	0	1	yes	1	no	0.79	0.05		1.2	11	7.6	489					4						
ZQ-10-83-CA-18	dravitic sst.	100	0	0	yes	2	no	0.67	0.05		0.6	9	2.5	1070					6						
ZQ-10-83-CA-20	siltstone	38	0	62	yes	1	no	0.59	0.15		3.4	7	10.3	483					24						
ZQ-10-83-CA-23	dravitic sst.	57	0	43	yes	4	no	0.77	0.14		11.9	9	4.0	5440					66						
ZQ-10-83-CA-25	silicified sst.	41	0	59	yes	4	no	0.68	0.03		0.2	11	1.7	3600					38						
ZQ-10-83-CA-27	silicified siltstone	31	11	58	yes	1	no	0.67	0.02		0.7	14	1.7	652					3						
ZK8D/345 [2]	dravitic sst.	15	14	71	yes	4	no	0.36	0.03	0.08	1.8	6	0.7	700	83	138	8	4	10	40	16	4	0.8	82	0.05
ZK8D/366	dravitic regolith?	37	0	63	yes	6	no	0.35	0.97	0.50	30	31	7.5	30000	3200	986	407	3	283	3200	1100	3	21.8	5654	0.02
HL8D/1717	dravitic sst.	32	67	1	yes	4	no	0.50	0.03	0.10	1.8	14	1.2	850	66	221	8	23	8	32	10	23	0.8	62	0.08
HL8D/1718	dravitic sst.	47	49	4	yes	4	no	0.39	0.04	0.15	2.8	9	0.9	600	99	280	7	19	8	38	13	19	0.9	74	0.07
MC6D/4840	dravitic sst.	59	0	41	yes	5	no	0.29	0.04	0.08	2.4	3	1.2	1200	124	232	11	2	5	60	30	2	0.8	126	0.03
MC6D/4846	dravitic sst.	20	0	80	yes	5	no	0.38	0.03	0.06	1.2	8	0.7	2100	80	124	5	3	3	46	17	3	0.5	91	0.03

Table 3 - Partial Microprobe Data: Maw Zone Data and Comparison with other published data. nc, not calculated; nd, not detected; 1 Demartin et al. (1991).

mineral	sample	condition	SiO_2 wt.%	ZrO_2 wt.%	U_3O_8 wt.%	ThO_2 wt.%	P_2O_5 wt.%	CaO wt.%	Al_2O_3 wt.%	Y_2O_3 wt.%	H_2O(c) wt.%	sum Y_2O_3+REO wt.%	sum REO wt.%	Y/(Y+REE) at.%
zircon	AM-84-13a	unaltered	32.8	65.0	0.1	0.0	nd	0.0		0.0	0.9	0.1	0.0	68.6
	AM-84-13b	unaltered	33.5	65.3	nd	nd	nd	nd		nd	0.0	0.1	0.1	nc
	AM-84-9a	mod. altered	30.0	62.1	nd	0.1	1.0	0.3		0.9	4.1	1.1	0.1	91.3
	AM-84-9b	mod. altered	27.9	60.4	0.2	0.1	2.0	0.5		1.6	5.7	1.7	0.1	96.1
	M-0101a	mod. altered	32.6	58.0	0.1	0.1	3.3	0.7	1.0	0.6	2.7	1.0	0.4	68.2
	AM-84-9c	altered	23.4	56.8	0.1	0.0	4.4	0.8	1.0	1.8	7.1	2.8	1.0	75.5
	M-0101b	altered	19.0	54.5	0.3	0.1	7.1	1.1	1.1	1.3	13.1	3.2	1.9	51.8
	AM-84-9d	altered	20.2	52.5	0.4	0.1	6.4	1.1		3.3	12.3	3.5	0.3	95.0
	M-0101c	altered edge	8.1	21.5	0.4	0.3	24.0	1.0	0.5	25.5	6.2	37.7	12.2	77.5
xenotime	Maw Zone	avg. (n=3)	0.6	nd	0.6	0.1	32.5	0.5	0.2	39.6	6.7	58.7	19.0	77.6
	Alpine[1]	avg. (n=82)	0.1	<0.01	1.0	0.4	35.1	0.1	<0.1	46.5	<0.1	63.9	17.4	81.8
detrital monazite	avg. (n=2)	unaltered	0.2	nd	2.5	6.8	25.0	2.0	nd	1.8	5.9	57.5	55.7	4.6

Table 4 - REO Data Recalculated to 100 percent REO: Maw Zone data and comparison with other published data. nc, not calculated; nd, not detected.

mineral	sample	condition	Y_2O_3 wt.%	Y/(Y+REE) at.%	Ce_2O_y/REO %	Eu_2O_y/REO %	Gd_2O_y/REO %	Tb_2O_y/REO %	Dy_2O_y/REO %	Ho_2O_y/REO %	Er_2O_y/REO %	Yb_2O_y/REO %	Lu_2O_y/REO %
zircon	AM-84-13a	unaltered	0.03	68.6									
	AM-84-13b	unaltered	<0.01	nc									
	AM-84-9a	mod. altered	0.94	91.3	0.0								
	AM-84-9b	mod. altered	1.61	96.1	83.3								
	M-0101a	mod. altered	0.57	68.2	16.3	<0.1	<0.1	2.3	<0.1	23.3	16.3	25.6	<0.1
	AM-84-9c	altered	1.83	75.5	1.0	<0.1	22.7	1.0	33.0	4.1	14.4	11.3	<0.1
	M-0101b	altered	1.26	51.8	5.7	<0.1	9.8	1.0	15.0	14.5	17.1	22.8	2.6
	AM-84-9d	altered	3.25	96.0	48.0								
	M-0101c	altered edge	25.47	77.5	0.5	<0.1	19.1	2.8	32.5	17.9	12.5	9.5	1.6
xenotime	Maw Zone	avg. (n=3)	39.64	77.6	0.3	nd	16.4	2.4	36.1	7.7	16.7	15.9	2.4
	Demartin et al. (1991)	avg. (n=82)	46.49	81.7	nd	0.23	10.57	3.44	29.38	6.00	22.34	24.18	3.13
	Jefford (1962)			68.5		nd	3.4	1.8	18.9	2.4	18.4	44.4	3.6
	Fleischer (1969)	avg. (n=54)		67.7		0.5	11.6	2.6	21.8	4.7	19.5	22.3	4.4
detrital monazite	Maw Zone	unaltered avg. (n=2)	1.85	4.6	51.2	nd	2.3	nd	1.1	0.7	<0.1	<0.1	<0.1



Plates 5 (upper left) and 6 (upper right) - Back-scattered electron images of altered zircon with xenotime rim at the Maw Zone. Q, quartz; Z, altered zircon; X, xenotime. Plate 7 (lower left) - Back-scattered electron image of altered zircon with a xenotime rim located at margin of fractured quartz grain in sandstone at Maw Zone. Q, quartz; Z, altered zircon; X, xenotime; T, dravite. Plate 8 (lower right)- Back-scattered electron image of discrete altered detrital zircon with a xenotime rim in sandstone at Maw Zone. Q, euhedral quartz; Z, altered zircon; X, xenotime; Tc, chrome dravite.

3. REE Mineral Chemistry

Rare earth element concentrations in xenotime (YPO_4) from the Maw Zone indicate a significant HREE enrichment. Similar HREE enrichments occur in uraninites and pitchblendes from unconformity-type uranium mineralization at Collins Bay, McArthur River, Key Lake and the Pine Creek Geosyncline (Figure 1; McClennan and Taylor, 1979; Fryer and Taylor, 1987) along with significant quantities of yttrium in uranium minerals from Key Lake and McArthur River (~3000 ppm). Concentrations of the REE in xenotime (170,000 ppm) are significantly higher than in the uranium minerals (2000 to 18,000 ppm) with the LREE/HREE ratios in xenotime and uraninite being about .01 and .10 to .16, respectively.

In contrast, illite containing varying amounts of goyasite and dravite from the alteration halos surrounding the unconformity-type uranium mineralization at McArthur River, Key Lake and Midwest Lake and dravite intergrown with xenotime in the alteration halo at the Maw Zone have a LREE enrichment with LREE/HREE ratios between 1.1 to 39.6 (Figure 1). In the unconformity-type uranium deposits, anomalous amounts of Sr in illite positively correlate with increasing goyasite content and concentration of LREE, suggesting that the LREE resides in the goyasite rather than the illite. Generally, illite with stoichiometric amounts of Sr has a flatter REE pattern with a LREE/HREE ratio slightly greater than 1.

4. Controls on REE and U Mobility in the Athabasca Basinal Fluids

The HREE and LREE contents in Maw Zone xenotime and dravite are similar to those in dravite, goyasite, clay minerals and uraninite from unconformity-type uranium deposits (Figure 1). These minerals have apparently formed from the same fluid processes, such that the

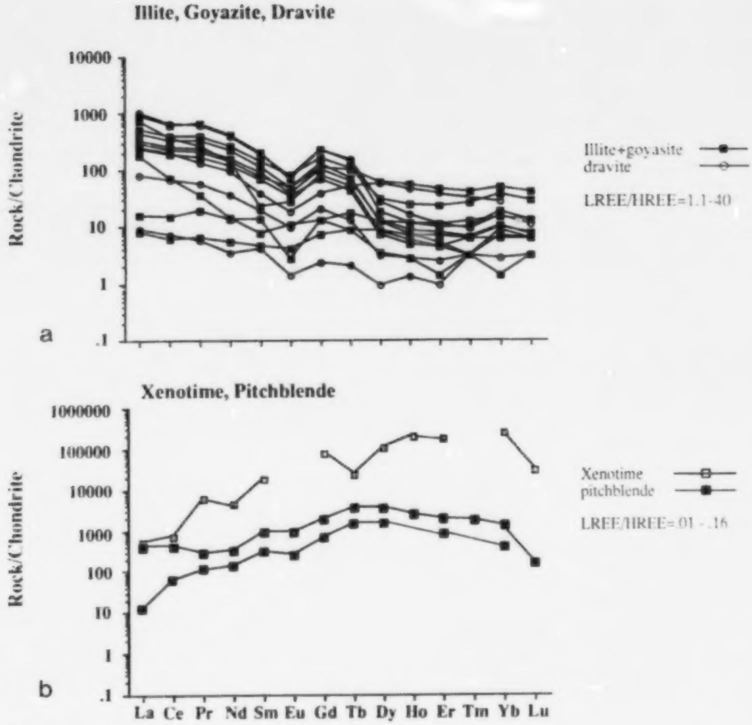


Figure 1 - Chondrite-normalized REE patterns for (a) illite, goyazite, dravite, and (b) xenotime and pitchblende minerals; LREE/HREE = (La + Ce + Nd)/(Dy + Er + Yb).

high REE concentrations and consistent REE patterns implies relatively homogeneous controls, for the transport and fractionation of REE, in areas of hydrothermal alteration throughout the basin. Similar Eu anomalies in all the minerals possibly suggests that the redox state was similar in the fluids responsible for hydrothermal alteration and uranium and HREE-yttrium mineralization.

Microthermometric analyses of fluid inclusions in gangue silicates and carbonates indicates that the Athabasca sandstones were altered by hot, highly saline, oxidizing basin fluids (Kotzer and Kyser, 1990b). These chloride-rich, basin brines are believed to have been the major source of the REE, uranium and precious and base metals. Experimental and empirical studies on the fluid transport of Au, U, PGE, yttrium and REE indicate they are highly mobile in chloride-rich brines similar to Athabasca Basin fluids (Michard, 1989; Wilde *et al.*, 1989; Wood, 1990). In the Athabasca Basin, anomalous REE, Y and U concentrations within the hydrothermal alteration haloes are compatible with high mobility of these elements in the chloride-dominated fluids which formed the Athabasca unconformity-type uranium deposits.

Studies of REE fractionation in various mineral groups (Burt, 1989; Mariano, 1989) along with empirical data on REE in hydrothermal fluids indicate that both Y and HREE tend to be mobile with, and concentrated in, xenotime and uranium relative to goyasite, dravite and illite, which tend to concentrate LREE. In the Athabasca unconformity-type uranium deposits, the HREE and Y enrichment in pitchblende and the LREE enrichment in the associated gangue and phosphate minerals are a consequence of the minerals which formed from the REE-enriched fluids. At the Maw Zone, formation of a Y-HREE-phosphate mineral (xenotime) possibly results from insignificant formation of pitchblende, which normally appears to deplete the basinal fluids of their HREE and Y content in the unconformity-type uranium deposits. Although whole-rock P_2O_5 contents of Maw Zone samples are, in general, comparable to background values for the Athabasca Group (Quirt, 1985), the conspicuous lack of goyazite in the Maw Zone area may indicate that this mineral was replaced by xenotime. However, individual samples contain phosphate contents up to 10 times background indicating that moderate phosphate enrichment is associated with the REE-Y mineralization process.

5. Sources of REE in the Athabasca Basin

Alteration of the Athabasca sandstones at the Maw Zone and McArthur River is due to mixing of fluids from the basin with those issuing from the basement along fault zones. Enrichments of REE in the fluids may have come from sources in either the underlying metamorphic rocks or the sandstones themselves. In sandstone environments, enrichments of heavy detrital minerals such as zircon can substantially alter the initial REE concentrations of the sediments, which generally tend to be low and LREE-dominated (McClennan, 1989). Garnets in the underlying garnet-bearing Wollaston Group gneisses and detrital zircons in the Athabasca sandstones, both enriched in HREE (McClennan, 1989), are sources for HREE in the basinal and basement fluids. At the Maw zone, textural relationships between the altered zircons and the HREE-enriched xenotime, coupled with the ability for the chloride-rich basinal brines to transport REE, indicates that the zircons were the more likely source for HREE in this area.

6. Phosphate Paragenesis in the Athabasca Basin

High mobility of phosphate is suggested from empirical and experimental studies on phosphate-clay assemblages which indicate that apatite dissolution and crandallite mineral precipitation can readily occur in neutral to low pH fluids from which illite and kaolinite can also form (Weaver and Wampler, 1972; Nriagu, 1976; Spotl, 1990). Mineralogical evidence from Athabasca Group sandstones indicates that the saline basin fluids were in equilibrium with illite and kaolinite, thus suggesting that phosphate mobility was possible. Phosphate mobility is also indicated by: early diagenetic apatite formation ca. 1650 to 1700 Ma (Cumming *et al.*, 1987); goyasite intergrown with illite and dravite ca. 1250 to 1500 Ma (Kotzer and Kyser, 1990a,b); and a very recent mixture of crandallite minerals with late-formed kaolinite (Hoeve and Quirt, 1984; Wilson, 1985; Kotzer and Kyser, 1990a). Broad correspondence of these phosphate ages to U-Pb, Rb-Sr, K-Ar and paleomagnetic ages on uraninite, diagenetic illite and hematite, which formed from regional high temperature fluid diagenesis (Kotzer and Kyser, 1990b), is compatible with the interpretation that the phosphate minerals also underwent continual destruction and reprecipitation by large scale fluid movements in the Athabasca sandstones.

7. Interpretation of Detrital and Diagenetic Dravite Compositions

Kotzer and Kyser (1990a) suggested that dravite formation in the alteration halos surrounding the unconformity-type uranium deposits resulted from mixing of a saline, oxidizing, boron-bearing, basin fluid with a reducing, Mg-rich basement fluid in the fault zones hosting the uranium deposits. The chemical and textural similarity of dravites from both the Maw Zone and unconformity-type uranium deposits and the alteration of detrital schorl-dravite indicate that the enclosing sandstones in both areas were altered by a basement-sandstone fluid mixing event involving a significant Mg component (Figure 2).

Slight chemical variations of the basement fluid are reflected by dravite compositions from the Maw Zone and unconformity-type uranium deposits. At the Maw Zone, formation of chrome dravite in pore spaces and in altered areas on the schorl-dravites, indicate elevated chromium concentration in the basement fluids. The sporadic occurrence of chrome dravite in close spatial association with a chrome-poor, alkali-free dravite implies that the basement fault zones penetrated rocks of distinctly different lithologies and that the basement fluids emanating from these fault zones had elevated chromium contents. The association of chrome dravites with ultramafic rocks (Dunn, 1977; King and Kerrich, 1989) suggests that along certain parts of the fault conduits the basement fluids possibly interacted with ultramafic rocks in the underlying Wollaston Group.

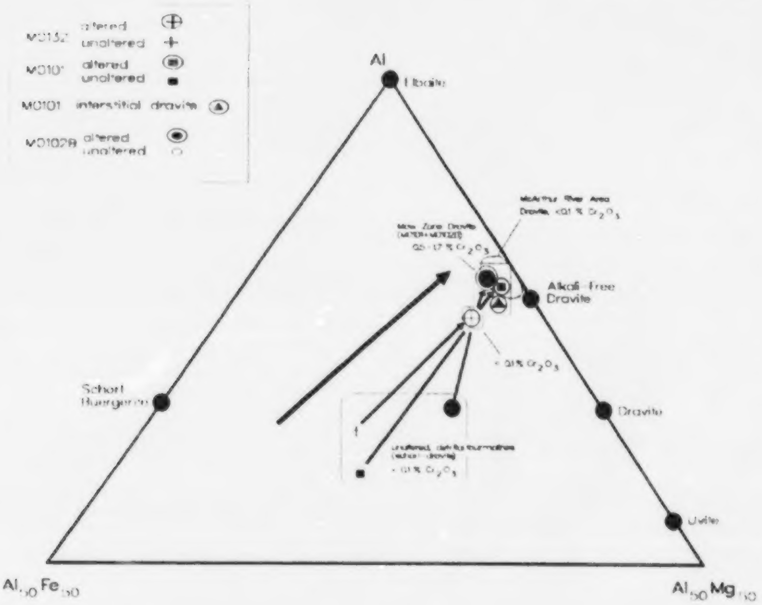


Figure 2 - Ternary plot (Fe-Mg-Al) of tourmaline compositions at McArthur River and the Maw Zone. Diagenetic dravite in pore spaces is an alkali-free dravite with variable amounts of chromium. Detrital tourmalines at the Maw Zone are schorl-dravites. Altered areas on detrital schorl-dravites have similar compositions to the interstitial alkali-free dravites. Arrows indicate trend of magnesium enrichment in the compositions of the tourmaline as a result of hydrothermal alteration.

8. Conclusions

In both the unconformity-type uranium deposits and the HREE-yttrium phosphate mineralization, similar spatial and textural associations between phosphate, silicate and clay minerals, common assemblages of euhedral quartz-dravite and similarities in the chemical compositions of dravite and REE concentrations in dravite, phosphates, clay minerals and uranium indicate a similar paragenesis and fluid history.

9. Acknowledgments

The senior author would like to express thanks to Agip Canada Ltd. for allowing outcrop mapping, trench sampling, and drill core logging and sampling during a site visit in 1984. Appreciation is due also to Saskatchewan Energy and Mines for allowing access in 1991 for sampling of the Maw Zone drill core stored in the La Ronge mineralized core collection. R. George of the University of Saskatchewan performed the

microprobe analyses. T.G.K and T.K.K. would like to thank D. MacDougall (Saskatchewan Energy and Mines) for providing them with Maw Zone samples.

10. References

Agip Canada Ltd. (1985): Assessment File 74H06-NW-0080, Sask. Energy and Mines.

Annesley, I.R. (1989): The Wollaston Group and its underlying Archean basement in Saskatchewan: An update; Sask. Resear. Counc., Publ. R-855-5-E-89, 86p.

Annesley, I.R. and Madore, C. (1988): The Wollaston Group and its underlying Archean basement in Saskatchewan: Preliminary report; *in* Summary of Investigations 1988, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 88-4, p54-60.

__________ (1989): The Wollaston Group and its underlying Archean basement in Saskatchewan: 1989 fieldwork and preliminary observations; *in* Summary of Investigations 1989, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 89-4, p87-91.

Burt, D.M. (1989): Compositional and phase relations among rare earth element minerals; *in* Geochemistry and Mineralogy of Rare Earth Elements, Miner. Soc. Am., Reviews in Mineral., v21, p259-302.

Cumming, G., Krstic, D. and Wilson, J. (1987): Age of the Athabasca Group, northern Alberta; Geol. Assoc. Can.-Mineral. Assoc. Can., Joint Annual Meeting, Progr. Abstr., v12, p35.

Deer, W., Howie, R., and Zussman, J. (1962): Rock Forming Minerals, Vol. 1 Ortho- and Ring-silicates; Longman, New York, 333p.

__________ (1966): An Introduction to the Rock Forming Minerals; Longman, New York, 528p.

__________ (1982): An Introduction to the Rock Forming Minerals; Halsted Press, J. Wiley and Sons, U.S.A., 528p.

Demartin, F., Pilati, T., Diella, V., Donzelli, S., Gentile, P. and Gramaccioli, C.M. (1991): The chemical composition of xenotime from fissures and pegmatites in the Alps; Can. Mineral., v29, p69-75.

Dunn, P.J. (1977): Chromium in dravite; Mineral. Mag., v41, p408-410.

Fleischer, M. (1989) Review of the distribution of the lanthanides in xenotime and related minerals; U.S. Geol. Surv., unpubl. report.

Fryer, B. and Taylor, R. (1987): Rare-earth element distributions in uraninites: Implications for ore genesis; Chem. Geol., v63, p101-108.

Hoeve, J. and Quirt, D.H. (1984): Mineralization and host-rock alteration in relation to clay mineral diagenesis and evolution of the middle-Proterozoic Athabasca Basin, northern Saskatchewan, Canada; Sask. Resear. Counc., Tech. Rep. No. 187, 187p.

Hoeve, J. and Sibbald, T.I.I. (1978): On the genesis of Rabbit Lake and other unconformity-type uranium deposits in northern Saskatchewan, Canada; Econ. Geol., v73, p1450-1473.

Hubregtse, J. and Sopuck, V. (1987): Alteration paragenesis of the Helikian sandstone and Aphebian basement of the Key Lake Mine and other uranium deposits in the eastern Athabasca Basin, Saskatchewan, Canada; IAEA, Technical Committee meeting on uranium resources and geology of North America, Saskatoon, 1987, abstr.

Jefford, G. (1962): Xenotime from Rayfield, northern Nigeria; Am. Mineral., v47, p1467-1473.

King, R.W. and Kerrich, R.W. (1989): Chromian-dravite associated with ultramafic hosted Archean lode gold deposits; Can. Mineral., v27, p419-426.

Kotzer, T.G. and Kyser, T.K. (1990a): The use of stable and radiogenic isotopes in the identification of fluids and processes associated with unconformity-type uranium deposits; *in* Beck, L.S. and Harper, C.T. (eds.), Modern Exploration Techniques, Sask. Geol. Soc., Spec. Publ. 10, p115-131.

__________ (1990b): Fluid History of the Athabasca Basin and its relation to Uranium Deposits; *in* Summary of Investigations 1990, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 90-4, p153-157.

__________ (1991): Retrograde alteration of clay minerals in uranium deposits: Radiation catalyzed or simply low-temperature exchange?; Chem. Geol. (Isotope Geosci. Sect.), v86, p307-321.

MacDougall, D.G. (1990): Rare earth element mineralization in the Athabasca Group - Maw Zone; *in* Summary of Investigations 1990, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 90-4, p103-105.

Mariano, A.N. (1989): Economic geology of rare earth minerals; *in* Geochemistry and Mineralogy of Rare Earth Elements, Miner. Soc. Am., Reviews in Mineral., v21, p309-334.

McClennan, S. (1989): Rare earth elements in sedimentary rocks: Influence of provenance and sedimentary processes; *in* Geochemistry and Mineralogy of Rare Earth Elements, Miner. Soc. Am., Reviews in Mineral., v21, p169-196.

McClennan, S and Taylor, S. (1979): Rare earth element mobility associated with uranium mineralization; Nature, v282, p247-249.

Michard, A. (1989): Rare earth element systematics in hydrothermal fluids; Geochim. Cosmochim. Acta, v53, p745-750.

Nriagu, J. (1976): Phosphate-clay mineral relations in soils and sediments; Can. J. Earth Sci., v13, p717-736.

Palache, C., Berman, H., and Frondel, C. (1951): Dana's System of Mineralogy; v2, 7th edition, John Wiley & Sons, New York, 1124p.

Quirt, D.H. (1985): Lithogeochemistry of the Athabasca Group: Summary of sandstone data; *in* Summary of Investigations 1985, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 85-4, p128-132.

Sibbald, T.I.I (1985): Geology and genesis of the Athabasca Basin uranium deposits; *in* Summary of Investigations 1985, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 85-4, p133-156.

Smith, D.G.W., De St. Jorre, L., Reed, S.J.B. and Long, J.V.P. (1991): Zonally metamictized and other zircons from Thor Lake, Northwest Territories; Can. Mineral., v29, p301-309.

Spotl, C. (1990): Authigenic aluminum phosphate-sulphates in the sandstones of the Mitterberg Formation, Northern Calcareous Alps, Austria; Sedimentology, v37, p837-845.

Weaver, C. and Wampler, J. (1972): The illite-phosphate association; Geochim. Cosmochim. Acta., v36, p1-13.

Wilde, A., Bloom, M. and Wall, V. (1989): Transport and deposition of gold, uranium and platinum group elements in unconformity-related uranium deposits; Econ. Geol., v84, p637-650.

Wilson, J.A. (1985): Crandallite group minerals in the Helikian Athabasca Group in Alberta, Canada; Can. J. Earth Sci., v22, p637-641.

Wilson, M. and Kyser, T.K. (1987): Stable isotope geochemistry of alteration associated with the Key Lake uranium deposit, Canada; Econ. Geol., v82, p1540-1557.

Wood, S. (1990): The aqueous geochemistry of the rare-earth elements and yttrium: 2. Theoretical predications of speciation in hydrothermal solutions to 350°C at saturation water vapor pressure; Chem. Geol., v88, p99-125.

The East Athabasca Mylonite Zone: An Archean Segment of the Snowbird Tectonic Zone in Northern Saskatchewan[1]

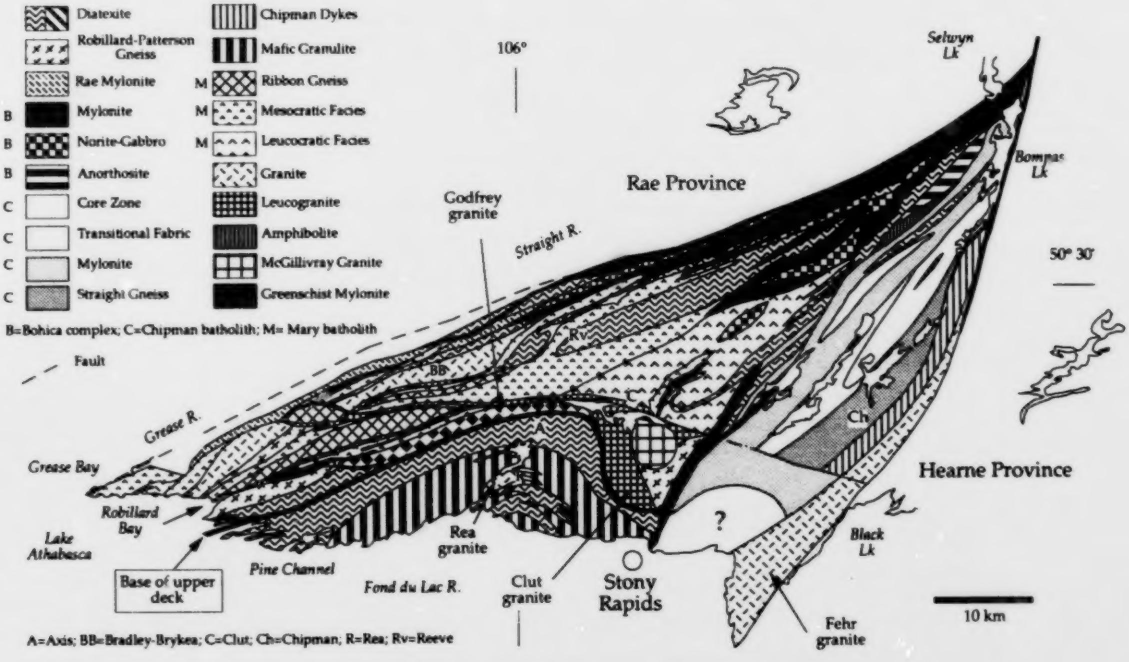
Simon Hanmer[2], Chris Kopf[3] and Mark Darrach[4]

Hanmer, S., Kopf, C. and Darrach, M. (1991): The east Athabasca Mylonite zone: An Archean segment of the Snowbird Tectonic Zone in northern Saskatchewan; *in* Summary of Investigations 1991, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 91-4.

The East Athabasca Mylonite Zone is a triangular segment of the Snowbird tectonic zone, characterized by a complex flow pattern. Polyphase deformation occurred under transitional to granulite facies conditions at ca. 3.2 and ca. 2.6 Ga.

The Snowbird tectonic zone, clearly delimited as a pronounced linear anomaly in the horizontal gravity gradient map of the Canadian Shield, is well exposed as a triangle of high grade mylonitic rocks (125 by 80 by 75 km) in the Stony Rapids area, northern Saskatchewan (Figure 1). The triangular structure (Tantato Domain of Gilboy, 1980; Slimmon, 1989) lies at the northeast end of the largest (300 by 100 km) of a chain of magnetically defined elliptical domains which may represent relatively stiff, crustal-scale 'boudins'(see Hanmer *et al*., 1991 for references).

Tantato Domain is divided into two structural decks (Figure 2; Hanmer *et al*., 1991). The fan-shaped lower deck is characterized by upright foliations and shallowly to moderately southwest plunging extension lineations. The upper deck is arcuate. Foliations parallel to the northwest-southeast-trending margin dip shallowly to the southwest and truncate the upright foliations of the lower deck, whereas those parallel to the northeast-southwest trending margin are upright. All extension lineations in the upper deck plunge shallowly, toward



Diatexite
Robillard-Patterson Gneiss
Rae Mylonite
B Mylonite
B Norite-Gabbro
B Anorthosite
C Core Zone
C Transitional Fabric
C Mylonite
C Straight Gneiss
Chipman Dykes
Mafic Granulite
M Ribbon Gneiss
M Mesocratic Facies
M Leucocratic Facies
Granite
Leucogranite
Amphibolite
McGillivray Granite
Greenschist Mylonite
B=Bohica complex; C=Chipman batholith; M= Mary batholith
Fault
106°
Selwyn Lk
Bompas Lk
Rae Province
Godfrey granite
Straight R.
50° 30'
BB
Rv
Ch
Grease R.
Grease Bay
Robillard Bay
Lake Athabasca
Base of upper deck
Pine Channel
Fond du Lac R.
Rea granite
Clut granite
Stony Rapids
?
Hearne Province
Black Lk
Fehr granite
10 km
A=Axis; BB=Bradley-Brykea; C=Clut; Ch=Chipman; R=Rea; Rv=Reeve

Figure 1 - Geology of East Athabasca Mylonite Zone.

---

(1) Funded by the Geological Survey of Canada and a support to C.Kopf from NSF grant to K. Collerson, Univ. of California, Santa Cruz.
(2) Continental Geoscience Division, Geological Survey of Canada.
(3) Earth Science Board, Univ. of California, Santa Cruz
(4) Department of Geology and Geography, Univ. Mass., Amherst, Massachusetts.

either the southwest or the northeast. Preliminary geochronological data (unpublished; U-Pb in zircon) have been obtained by R. Parrish (Geochronology Laboratory, Geological Survey of Canada). In order to more clearly convey the structural nature of the Snowbird tectonic zone in the map area, we suggest that the name East Athabasca Mylonite Zone (referred to as the Mylonite Zone for brevity) should replace the non-genetic term 'Tantato Domain'.

1. Upper Deck

a) Pine Channel Diatexite

The Pine Channel diatexite, 4 km thick, is white to tan and comprises a banded, leucocratic, quartzofeldspathic matrix with an highly variable content of lilac to delicate pink garnets, orthopyroxene, graphite and sillimanite. Mylonitic to ultramylonitic ribbon fabrics are penetratively developed throughout the Pine Channel diatexite, except near Pine Channel. There, kilometric lenses of often highly irregular, folded migmatite with cross-cutting garnet-orthopyroxene granitoid sheets form a stack of steeply southeast-dipping 'horses', bounded above and below by their highly attenuated ultramylonitized equivalents.

b) Fond du Lac Mafic Granulites

The Fond du Lac mafic granulites are a 5 km thick stack of sheets (< 2 km thick), essentially orthopyroxene-plagioclase-magnetite ±garnet in composition, separated from one another by thin, but often mappable, horizons of diatexite ribbon mylonite. South and east of Axis Lake, the mafic granulites preserve isotropic coarse ophitic gabbroic/noritic igneous textures over a thickness of more than a kilometre, which grade into annealed mylonites.

2. Lower Deck

a) Bohica Complex

Very coarse, primary ophitic textures to strongly flattened, flaser fabrics are locally preserved in norite, gabbro and diorite, especially in the central part of the Mylonite Zone. The coarse igneous textures can be followed progressively into garnet-two pyroxene to garnet-hornblende (±pyroxene) mylonites. Two large bodies of anorthosite and gabbroic anorthosite ultramylonite, locally with splendid ribbon fabrics, locally strongly annealed, outcrop on the southeast side of the complex.

b) Chipman Batholith

In the tonalitic Chipman batholith (pre-ca. 3.2 Ga), Hanmer *et al.* (1991) identified a coarse-grained, low strain central core, flanked to the southeast by strongly annealed tonalitic straight gneisses (mylonites), and to the northwest by clinopyroxene-bearing tonalite ribbon mylonites and ultramylonites. Widespread xenoliths and rafts represent the remains of a dismembered layered mafic-anorthositic-ultramafic complex. An important swarm of garnet-hornblende and garnet-clinopyroxene Chipman dykes was emplaced throughout the batholith. However, east of a line through Chipman Lake, and spatially coincident with the strongly annealed tonalitic straight gneisses, the dykes are locally so common that

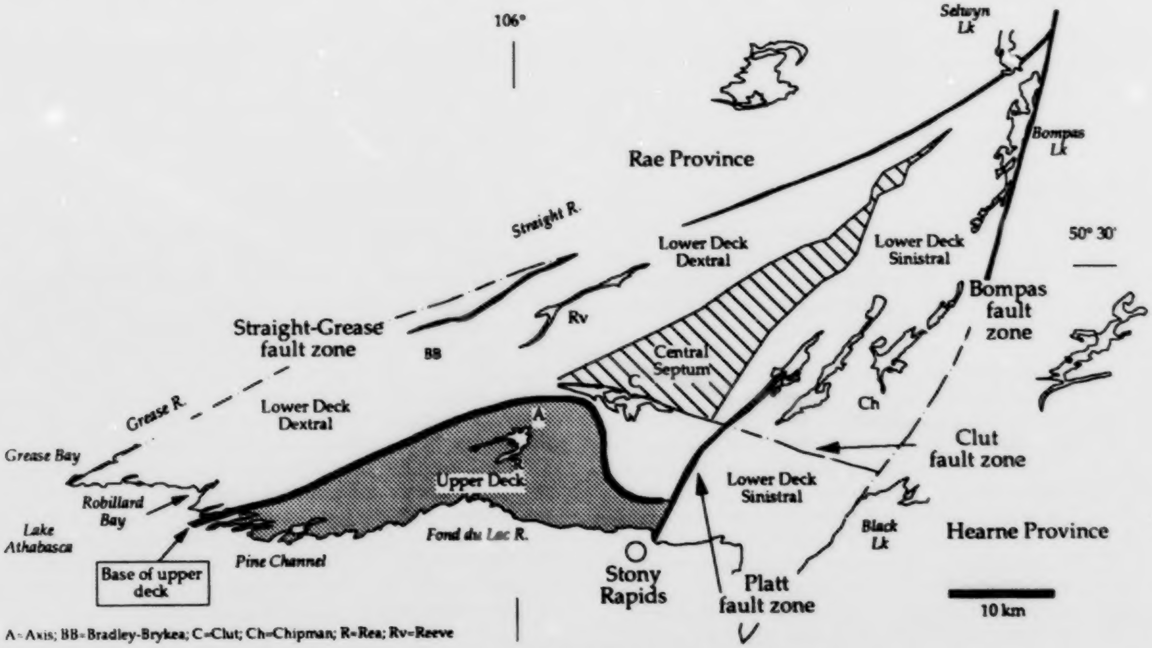

106°
Selwyn Lk
Bompas Lk
Rae Province
Straight R.
Lower Deck Dextral
Lower Deck Sinistral
50° 30'
Bompas fault zone
Straight-Grease fault zone
Rv
BB
Central Septum
C
Grease R.
Lower Deck Dextral
A
Ch
Clut fault zone
Grease Bay
Upper Deck
Lower Deck Sinistral
Robillard Bay
Lake Athabasca
Fond du Lac R.
Black Lk
Hearne Province
Base of upper deck
Pine Channel
Stony Rapids
Platt fault zone
10 km
A=Axis; BB=Bradley-Brykea; C=Clut; Ch=Chipman; R=Rea; Rv=Reeve

Figure 2 - Structural framework of East Athabasca Mylonite Zone.

they may form 70 (or more) percent of the map unit in belts up to 1.5 km thick. The dykes show a complete range of deformation, from early transposed components of the tonalite gneissosity, to later cross-cutting, isotropic dykes with pristine igneous contacts. The earlier, strongly foliated dykes and the tonalitic straight gneisses are crosscut by an array of 1 to 10 m thick leucogranite veins (Hanmer *et al.* 1991; ca. 3.2 Ga.).

c) Mary Batholith

Hanmer *et al.* (1991) first distinguished the Mary batholith as a compositionally monotonous, strongly garnetiferous meta-granite to meta-granodiorite. The garnets occur in two distinctive microstructural settings: as 2 to 5 mm single grains rimmed by hornblende, apparently after pyroxene (core-type), and as small (<1mm) grains arranged in rims, one grain thick, mantling aggregates of hornblende after clinopyroxene (rim-type). The abundance of mafic minerals lends a mesocratic aspect to the Mary granitoids west of Mary, hence the term western mesocratic facies (ca. 2.6 Ga.).

By contrast, granitoid rocks in the eastern part of the Mary batholith are distinctly leucocratic (eastern leucocratic facies; ca. 2.6 Ga.), yet contain the same mineral assemblages and garnet microstructures just described. Locally, there are map-scale panels of the mesocratic Mary material within the leucocratic granitoids. Elsewhere, the garnet-hornblende-clinopyroxene content of the leucocratic granitoid is highly variable, both between and within individual outcrops. In the latter case the transitions are always gradational. This strong mineralogical variation is an hallmark of the Mary batholith. The presence of a map-scale raft of garnet-pyroxene layered metagabbro with well preserved cumulate layering, plus abundant evidence of the mechanical disaggregation of fist to football sized garnet-pyroxene aggregates within the host granitoid, suggest to us that part of the mineralogical heterogeneity in the Mary batholith may reflect variation in the xenocryst population at all scales.

d) Other Granites

Discrete, map-scale plutons and sheets of generally leucocratic granite occur throughout the lower deck of the Mylonite Zone. They probably do not represent a related suite of granites and are only grouped here for clarity of presentation.

The Godfrey granite was described by Hanmer *et al.* (1991) as a coarse-grained, garnet-pyroxene metagranite, lining the base of the upper deck in the Clut Lakes area. It may closely resemble the meso- to leucocratic facies of the Mary batholith. We cannot exclude the possibility that this granite is related to the Mary batholith. The Brykea granite is a strikingly salmon pink granite ultramylonite with abundant, 1 mm cherry red garnets set in a fine-grained, sugary annealed matrix with variable proportions of small feldspar porphyroclasts. The Bradley granite is a splendid ribbon ultramylonite derived at the expense of a very coarse grained protolith. The Fehr granite (ca. 2.6 Ga.) outcrops along the eastern margin of the Mylonite Zone. To the southwest, it is generally isotropic and extremely coarse grained, with the exception of a few, local, discrete shear zones with upright fabrics, trending subparallel to the regional strike. To the southeast, the granite is charged with abundant amoeboid clots of coarse, polycrystalline hornblende ±orthopyroxene, visibly derived by the disaggregation of larger, metre-size xenoliths, reminiscent of the xenocrystic mafic phases of the Mary batholith.

The granite is cut by a swarm of strike-parallel mafic dykes, one to tens of metres wide, which may locally account for 20 percent of the Fehr granite. Some of the dykes carry abundant 2 to 4 mm garnets, commonly replaced by plagioclase. Although its contact with the Chipman batholith is abrupt, veins of Fehr granite cut across both the batholith structure and the Chipman dykes. Accordingly, there are at least two dyke swarms in the eastern part of the Mylonite Zone.

The Clut granite (ca. 2.62 Ga.) is a voluminous biotite-hornblende leucogranitic protomylonite which occupies the Clut Lakes area and bounds the internally unfoliated, late-syntectonic megacrystic hornblende-biotite McGillivray granite (Hanmer *et al.*, 1991; ca. 2.62 Ga.) on its western and northern sides. Note that the McGillivray granite intruded the already deformed Clut granite (Hanmer *et al.*, 1991).

Along strike to the west and south of the Clut Lakes granite is a suite of structurally controlled biotite-hornblende leucogranite sheets, each one up to several hundreds of metres thick, or more. They lie preferentially within the belts of porphyroclastic mylonite which enclose the map-scale lozenges of ribbon gneiss and Reeve diatexite. We refer to them as the Melby-Turnbull granites. Although the isotropic to poorly foliated parts of the granite sheets have abrupt intrusive contacts and contain misoriented xenoliths of the adjacent porphyroclastic mylonites, they are themselves locally to extensively mylonitised. We suggest that they were emplaced during the waning stages of mylonitisation.

e) Robillard-Patterson Gneiss

Two map-scale slices of a white granitoid gneiss, with included rafts of rhythmically layered garnet-sillimanite meta-(psammo)pelite, occur within the lower deck, east of Robillard Bay and west of Patterson Creek. The granitoid component, is well banded, coarsely rodded and irregularly folded. At the grain-scale, the rock is coarsely recrystallised and generally lacks a shape fabric. Pegmatite veins in a variety of strain states are common and may crosscut both banding and folding in the gneiss. In the eastern part of the Patterson Creek slice, the irregular, intimate mixture of granitic and metasedimentary components is distinctly migmatitic; in the western part, the very coarse-grained granitoid intrudes and includes a large mass of isotropic gabbro. We suggest that the fabrics in the Robillard-Patterson gneiss may be older than the deformation and metamorphism of the adjacent Mylonite Zone rocks.

3. Structure and Metamorphism

Foliations within the fan-shaped lower deck (Figure 2) are generally steeply dipping and concordant with the Mylonite Zone boundaries, except in the southwest, along the shore of Lake Athabasca. Here, the foliation is commonly very shallow dipping and highly variable in orientation. Nonetheless, extension lineations tend to plunge moderately to shallowly southwestward, irrespective of foliation attitude.

The distribution, development and preservation of high strain fabrics in the lower deck are complex. The principal low strain zone is the Central Septum (Figure 2) of heterogeneously deformed Bohica complex mafic rocks and leucocratic Mary batholith granites which extends northeast from the Clut Lakes. It is flanked on either side by extremely attenuated Reeve diatexite, Chipman batholith and Bohica complex granulite facies ribbon mylonites.

Outside of the relatively low strain zones, the rocks of the lower deck are highly attenuated mylonites or straight gneisses, with the exception of the Robillard-Patterson gneiss. To the northwest of the Central Septum, garnet-hornblende-pyroxene (±sillimanite) ribbon mylonites are developed from Reeve diatexite, Bohica complex and Mary batholith lithologies. Within 2 km of the northwest margin of the Mylonite Zone, mylonites are so fine grained that the ribbon structure is difficult to discern in the field. Biotite-chlorite-bearing greenschist mylonites derived from metapelitic and leucogranite sheets form a 100 m wide belt at the northwest edge of the Mylonite Zone. To the southeast of the Central Septum, garnet-two pyroxene ribbon mylonites are developed from Reeve diatexite, anorthosite and mafic rocks of the Bohica complex and from the Chipman batholith. A narrow belt, less than a kilometre thick, of garnet-hornblende ribbon mylonites is developed along the eastern margin of the Mylonite Zone in rocks of the Chipman batholith and the northern part of the Fehr granite. These pass eastward into a very narrow zone, about 100 m thick, of biotite-chlorite mylonites derived from metapelitic rocks and the western margin of the Bompas granite.

Within the upper deck, the strike of the foliation in the penetratively developed orthopyroxene-plagioclase, garnet-orthopyroxene and sillimanite-orthopyroxene ribbon mylonites is concordant with respect to the deck margin, whereas the extension lineation is everywhere oriented northwest-southeast. Although the base of the lower deck coincides with a lithological boundary in the east, in the west it lies within the Pine Channel diatexite (Figure 3). The boundary is everywhere the convex limit of penetrative, completely transposed ribbon mylonites. Within the adjacent lower deck, the Pine Channel diatexite is a ribbon mylonite with abundant cross-cutting and open to tight folded granite veins. At the boundary of the upper deck, the granite sheets are totally transposed and the ribbon fabric becomes even more extremely attenuated.

4. Acknowledgments

We are indebted to Melanie Kells, Brent Loney, Mike Peshko and Dave Ross for their enthusiasm and unflagging contribution to this summer's fieldwork, regardless of bush, terrain or weather. CK acknowledges support from NSF grant EAR 9105284 to K. Collerson at U.C.S.C.

5. References

Gilboy, C. F. (1980): Bedrock compilation geology: Stony Rapids area (NTS 74-P), Preliminary Geological Map; *in* Summary of Investigations 1980, Saskatchewan Geological Survey, Sask. Miner. Resour., Misc. Rep. 80-4, scale 1: 250 000.

Hanmer, S., Ji, S., Darrach, M. and Kopf, C. (1991): Tantato domain, northern Saskatchewan: a segment of the Snowbird tectonic zone; *in* Current Research Part A, Geol. Surv. Can., Pap. 91-1A, p121-133.

Slimmon, W.L. (1989): Bedrock compilation geology: Fond du Lac (NTS 74-O); Sask. Energy Mines, Rep. 247.

D. Southern Projects Under the Canada-Saskatchewan Partnership Agreement on Mineral Development 1990-95, in Year 1991-92

Review of Industrial Minerals Investigations 1991-92[1]

P. Guliov and M. Gent

Guliov, P. and Gent, M. (1991): Review of industrial minerals investigations 1991-92; *in* Summary of Investigations 1991, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 91-4.

Industrial mineral investigations by the geological survey provide a foundation for detailed evaluation by industry and a catalyst for development and diversification. Most individual projects arise from perceived long and short-term mineral-based developmental opportunities for the province, from the recognition of information gaps, and from the need for data enhancement.

Most projects undertaken by the branch during the fiscal year were funded under the Canada-Saskatchewan Partnership Agreement on Mineral Development (PAMD), but some which may be related to PAMD projects, are funded solely by the province. The 1991-92 program includes the following projects:

1. Indicator Mineral Sampling and Analyses
2. Clay Studies
3. Industrial Mineral Sampling
4. Coalfield Modelling
5. Southern Gravity Data Compilation
6. Industrial Minerals Computerized Data Base Development
7. Computerized Index of Geophysical Data
8. Brine Digital Data Base Compilation

1. Indicator Mineral Sampling and Analyses

Exploration for diamondiferous kimberlites or lamproites remains the single, largest mineral exploration activity within the Phanerozoic Basin of Saskatchewan. Limited regional sampling of glaciofluvial gravels in the Big Muddy Valley area during 1990 obtained some significant results. Indicator minerals were found that suggest the presence of kimberlitic rocks and ruby-bearing intrusives. These and other preliminary results of investigations for precious gems in Saskatchewan are described in a report to be released shortly.

Sampling during the summer of 1991 expanded the regional coverage to include all of the south-central and eastern areas of Saskatchewan. During the sampling program, significant quantities of kimberlite indicator minerals of unusually large size were found in the Val Marie area. These minerals occur in the glacial till, and suggest that kimberlitic intrusives are in close proximity.

2. Clay Studies

Clay resources of Saskatchewan comprise swelling and non-swelling bentonites, and earthenware, stoneware, ball, refractory, kaolin and bloating clays. Early research and evaluation of kaolinitic clays in the province concentrated largely on their ceramic properties and applications. Bentonites were largely evaluated for their application in foundries and iron ore pelletizing and as drilling mud and clarifying agents for oils. More recently Saskatchewan clays have been considered as potential candidates for more specialized products such as fillers, carrier media, surface-treated products, raw materials for secondary products such as zeolites and a variety of other specialty products. Industrial use opportunities for Saskatchewan clays are substantial, but development is notably restrained.

The clay resource data base maintained by the Saskatchewan Geological Survey is highly varied in terms of its scope and quality and consists of numerous published and unpublished documents as well as assessment reports dating from the early part of the century. The data base is the starting point for the clay project whose principal objective is to stimulate the industrial development of the various resources. This objective is addressed through several subsidiary objectives as follows:

1. compile and evaluate the technical and economic data base and determine the nature of required data enhancement in light of modern clay technology and industrial applications;
2. priorize the clay resources for their developmental and technical-economic potential;
3. enhance the technical and economic data base as required through drilling, sampling, analyses and other investigations;
4. produce a report or series of reports dealing with the technical and economic aspects of Saskatchewan clay resources to be distributed among potential developers.

The project has three phases which will require three to four years to complete. The first phase, to be completed in the current fiscal year, includes the following:

1. data compilation and evaluation;
2. planning of field and analytical work for the enhancement of the clay data base to the desired level;

---

(1) These are projects funded under the Saskatchewan component of the Canada-Saskatchewan Partnership Agreement on Mineral Development 1990-95.

3. limited field investigations, sampling and analytical work on kaolinitic clays and bentonites;
4. commencement of clay resource maps; and
5. preliminary report preparation.

3. Industrial Minerals Sampling

This project is a continuation of the sampling project begun in 1990-91. The principal objectives are to sample and conduct physical and chemical analyses of minerals which are regarded as having significant developmental potential. In most cases several hundred kilograms of material are collected for testing and future reference. It is planned that representative samples would be made available to prospective developers for evaluation. An open-file report on the preliminary results of these investigations should be released shortly.

Sampling this fiscal year was restricted to the silica sand deposits of the Ordovician Winnipeg Formation in the Hanson Lake area. The section was channel-sampled at about one metre intervals over a thickness of 4.6 m. The sand is being tested in a similar manner to the previous year's silica sand samples. Among other possibilities the material appears to have potential as a hydraulic fracturing sand.

4. Coalfield Modelling

The coalfield modelling project continues work conducted in previous years under the Canada-Saskatchewan Fossil Fuels Technology Development Agreement. The project, started in 1988, has since been transferred to the MDA and PAMD, and is a co-operative effort between the Geological Survey of Canada and Saskatchewan Energy and Mines.

The number of coal exploration boreholes in southern Saskatchewan has tripled since the last major resource assessment in 1978. Computerized geological modeling techniques have also advanced greatly since then. Thus, with the need for a more accurate and practical definition of coal resources the present project was developed and consists of five principal elements:

1. capture, interpretation, verification and digitization of data including geophysical logs, drillers' logs, cuttings descriptions, and analytical data;
2. geological correlation to develop the necessary stratigraphic framework;
3. computer modelling of the coalfields;
4. integration of other data into the system including ash and sulphur content of the coal, lithology of the interbeds, land use data, topographic information and mining data; and
5. resource analysis and reporting.

Saskatchewan Energy and Mines was involved primarily in parts 1 and 2 (above) through the contractual services of a coal geologist. During the present fiscal year the main thrust of the project is on:

1. the completion of the final report on the Estevan coalfield modelling and resources definition;
2. the integration of additional borehole data into the system; and
3. the completion of the Wood Mountain-Willow Bunch digital model.

The completed model will be of value in coal resource conservation and allocation, policy decisions, recoverable reserve calculations, industrial development and land use planning, projected security of energy supply estimates, environmental considerations, and other applications.

5. Southern Gravity Data Compilation

This project was initiated in 1990 to optimize the use of gravity data records stored with Saskatchewan Energy and Mines. The data are in the process of being digitized, leveled and corrected to the WGS 84 ellipsoidal model. It is anticipated that this data set will constitute the largest detailed gravity data base in Canada and will be of significance to most if not all exploration activities in the region, including those for petroleum, potash, diamonds, and base and precious metals.

6. Industrial Minerals Computerized Data Base Development

The industrial minerals file is a large, complex and highly varied, commodity-based manual data base. Its use for a wide variety of purposes relating areal, local and topical investigations and queries is complex and time-consuming. Much of this data base is highly amenable to development in digital form.

This project addresses the need for computerizing the industrial minerals data base. The principal objective is to develop a versatile, cost effective computerized system that is easy to use and maintain. It is a large undertaking and will require several years to accomplish.

Work during the current fiscal year focused on overall planning of the system and the design and testing of data entry forms. To date forms have been designed for water analyses, gravity data, silica sand, and disposition data for coal, potash, quarriable minerals, alkali minerals and other minerals under the mineral disposition regulations. Data entry for system testing has been done for water analyses, gravity surveys and silica sand.

7. Computerized Index of Geophysical Data

The Saskatchewan Geological Survey maintains a large geophysical data base pertaining to the sedimentary basin of the province. The data base comprises seismic contour maps, gravity and magnetic survey maps, seismic shotpoint location maps, flowing seismic shot-hole data , structure test hole information, other miscellaneous maps, and reports relating to geophysical surveys. A file of current geophysical survey activity in the province is also maintained.

Computerized indexing of these data subsets has continued as a summer student project since 1988 and is nearly complete. The digitized file provides a means of rapid searches for available data in any specified area and it may also be presented in catalog format customized to individual needs. The file will save considerable data-searching time and will be of benefit to the petroleum and mineral exploration industries.

8. Brine Digital Data Base Compilation

Brines constitute one of the principle sources of industrial minerals used in the chemical industries of the world. To date most of the brines used in Saskatchewan have been produced by induced dissolution of potash and rock salt in the Prairie Evaporite Formation. The natural brine resources of Saskatchewan represent a vast, rich source of calcium and magnesium. The Potash Corporation of Saskatchewan is the only company in the province presently producing a natural brine from its Cory mine.

A joint Canada-Saskatchewan project was initiated in 1990 to compile a digital data set of borehole water quality analyses. A Sub-Duperow digital data base containing over 1200 borehole water analyses was released in April 1991 and an open file report is scheduled for release shortly. The final data base version containing approximately 14,000 analyses is scheduled for release by October of this year.

Industry-Government Aeromagnetic Survey of Southern Saskatchewan[1]

D. Teskey[2], M.R. Gent and P.E. Stone[2]

Teskey, D., Gent, M.R. and Stone, P.E. (1991): Industry-government aeromagnetic survey of southern Saskatchewan; *in* Summary of Investigations 1991, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 91-4.

The Geological Survey of Canada (GSC), and Saskatchewan Energy and Mines (SEM) with members of the mineral exploration industry initiated a program to complete the regional aeromagnetic coverage of Saskatchewan in four years. The area flown in 1991 is outlined in Figure 1.

Specifications as selected by the industry-GSC-SEM technical committee are 150 m terrain clearance at 800 m line separation in an east-west direction. The sensitivity of the system is better than 0.1 nT and positioning by GPS is better than 25 m. The survey parameters will ensure that the data is useful for kimberlite, potash and hydrocarbon exploration as well as detailed mapping of the sub-Phanerozoic basement. Participation in the program is on a yearly basis with participating companies receiving three years exclusive use of the data before release to the public. More information can be obtained from:

Malcolm Gent, SEM (306) 787-2646,
Dennis Teskey, GSC (613) 992-9763.

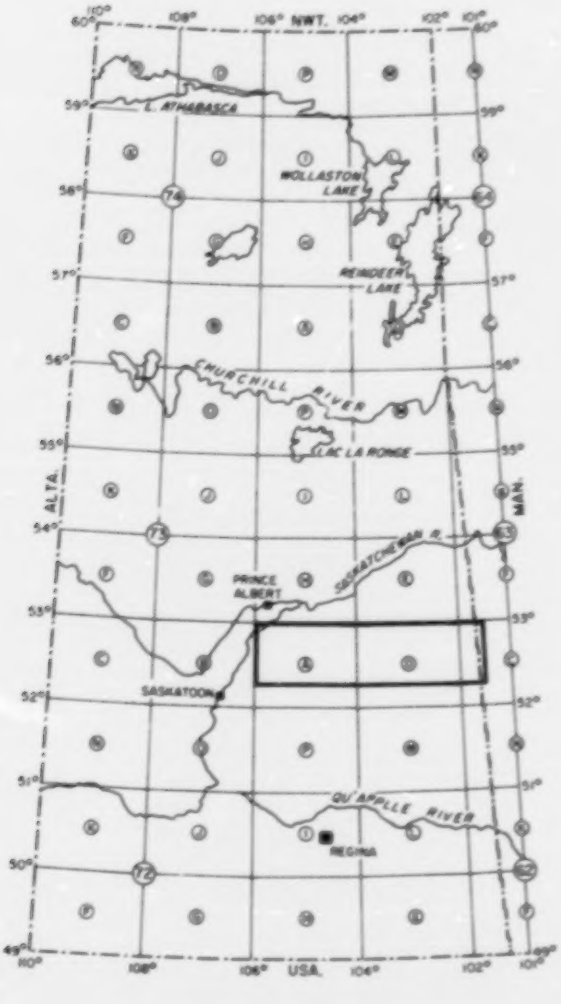

110°
108°
106° NWT.
104°
102°
101°
60°
59°
L. ATHABASCA
WOLLASTON LAKE
58°
74
64
57°
REINDEER LAKE
56°
CHURCHILL RIVER
55°
LAC LA RONGE
ALTA.
MAN.
54°
73
63
SASKATCHEWAN R.
PRINCE ALBERT
53°
SASKATOON
52°
51°
QU'APPELLE RIVER
REGINA
50°
72
62
49°
USA.

Figure 1 - NTS map of Saskatchewan showing location of the area surveyed in 1991.

---

(1) Saskatchewan Project A.221, from which this report derives, was jointly funded in 1991 under the Canada-Saskatchewan Partnership Agreement on Mineral Development 1990-95, and by Cameco Corporation, Uranerz Exploration and Mining Limited and Rio Algom Exploration Inc.
(2) Geological Survey of Canada, 601 Booth St., Ottawa, Ontario, K1A 0E8.

E. Other Southern Studies

Projects Supported by Petroleum Geology Branch in 1991

No.	Researcher	Title	Area	Status
PG861	Haidl, F.M.	Geology of the Silurian Interlake Formation, Saskatchewan	Townships 1 to 60 inclusive	Ongoing in Oct. 1991
PG862	Gilboy, C.F.	Geology of the Uper Colorado Group and the Milk River Formation (Upper Cretaceous), Southwestern Saskatchewan	Ranges 21 to 29W3 inclusive; Townships 1 to 20 inclusive	Ongoing in Oct. 1991
PG871	Christoper, J.E.	Geology of the Mannville Group (Lower Cretaceous), Saskatchewan	Townships 1 to 100 inclusive	Contract. Writing up Oct. 1991
PG872	Kreis, L.K.	Stratigraphy of the Jurassic System in the Wapella-Moosomin Area, Southeastern Saskatchewan	Ranges 30W1 to 2W2 inclusive; Townships 12 to 17 inclusive	Study completed March 1990 as Master's Thesis; successfully defended April 1990; report published April 1991
PG901	Gilboy, C.F.	Regional Cross-sections of Strata above the Mannville Group (Lower Cretaceous) in Saskatchewan	Townships 1 to 25 inclusive	Ongoing Oct. 1991
PG902	Leckie, D.A., Gilboy, C.F. *et al.*	Colorado/Alberta Group Strata of the Western Canada Sedimentary Basin	Townships 1 to 65 inclusive (in Saskatchewan)	Completed and presented at 'Basin Perspectives' Meeting, Calgary, May 1990; expanded for incorporation into 'Geological Atlas of the Western Canada Sedimentary Basin' in late 1991
PG903	Slind, O.L., Paterson, D.F. *et al.*	Middle Cambrian-Lower Ordovician Strata of the Western Canada Sedimentary Basin	Townships 1 to 70 inclusive (in Saskatchewan)	Completed and presented at 'Basin Perspectives' Meeting, Calgary, May 1990; expanded for incorporation into 'Geological Atlas of the Western Canada Sedimentary Basin' in late 1991
PG904	Norford, B.S., Haidl, F.M., Paterson, D.F., *et al.*	Middle Ordovician-Silurian Strata of the Western Canada Sedimentary Basin	Townships 1 to 66 inclusive (in Saskatchewan)	Completed and presented at 'Basin Perspectives' Meeting, Calgary, May 1990; expanded for incorporation into 'Geological Atlas of the Western Canada Sedimentary Basin' in late 1991
PG906	Kreis, L.K. (and Gent, M.)	Subsurface Brines in Southern Saskatchewan	Townships 1 to 80 inclusive	Poster display at S.E.M. Open House 1990. Open File Report in prep., Oct. 1991; paper prepared for presentation and publishing at 6th Intnat. Williston Basin Symp., Oct. 1991
*PG911	Haidl, F.M.	Note on the Ordovician-Silurian Boundary in Southeastern Saskatchewan	Townships 1 to 20 inclusive; Ranges 30W1 to 20W2 inclusive	Paper for Summary of Investigations, Nov. 1991
*PG912	Gilboy, C.F. and Nambudiri, E.M.V.	Note on Preliminary Palynological Investigations of Upper Cretaceous Strata, Southwestern Saskatchewan	Townships 2 to 16 inclusive; Ranges 28W3 to 30W3 inclusive	Paper for Summary of Investigations, Nov. 1991
UR891	Yurkowski, M.	Pore Geometry Reservoir Model Study for the Upper Winnipegosis Member (Devonian) in Southwest Saskatchewan	Ranges 30W1 to 12W2 inclusive Townships 1 to 10 inclusive	Ongoing Master's Study at University of Regina in Oct. 1991; progress report May 1991

*Projects are reported on in this volume.

Note on the Ordovician-Silurian Boundary in Southeastern Saskatchewan

F.M. Haidl

Haidl, F.M. (1991): Note on the Ordovician-Silurian boundary in southeastern Saskatchewan; *in* Summary of Investigations 1991, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 91-4.

1. Stratigraphy

Uppermost Ordovician and lower Silurian strata (Stonewall and Lower Interlake, Figure 1) in the Williston Basin are characterized by carbonate-evaporite cyclic sedimentation (Kendall, 1976; Osadetz and Haidl, 1989; Norford *et al.*, in press). Each cycle, where completely developed, consists of four main units: 1) fossiliferous, commonly burrowed, dolomite, 2) dolomite mudstone, 3) anhydrite, and 4) argillaceous dolomite and dolomitic shale marker beds which commonly contain quartz grains.

The fossiliferous unit is typically a wackestone. Laminated intervals are abundant in the dolomite mudstone unit and are commonly associated with stromatolites, particularly in the Interlake (Johnson and Lescinsky, 1986). In Saskatchewan the cycles include anhydrite beds in the Stonewall Formation only in the southeastern corner of the province (south of Township 12, east of Range 23W2). Evaporites in the Lower Interlake are restricted to the central portion of the Williston Basin in North Dakota.

In the subsurface, the cycles are best defined by the widespread argillaceous, locally arenaceous, dolomite marker beds which have a high gamma ray log response. The mechanism(s) responsible for the deposition of these marker beds is (are) not well understood. The beds may represent the regressive phases of transgressive-regressive cycles related to glacio-eustatic sea level fluctuations (Johnson and Lescinsky, 1986).

The base and top of the Stonewall Formation are defined by two of these marker beds (Figure 1)(Porter and Fuller, 1959; Kendall, 1976). Porter and Fuller (1959) also named a number of marker beds within formations. Of particular interest is the t horizon within the upper half of the Stonewall. Kendall (1976) expanded the t horizon to include an upper and lower t horizon in several wells where two closely-spaced marker beds occur at this stratigraphic interval (e.g., Figure 1). In the type well of Porter and Fuller (1959), Imperial Hummingbird 6-13-2-19W2, the t horizon corresponds to Kendall's upper t horizon. The more argillaceous lower t horizon is most commonly picked, however, when only one t horizon is named in wells with two marker beds in this zone (e.g., Brindle, 1960; Kendall, 1976; Norford *et al.*, in press).

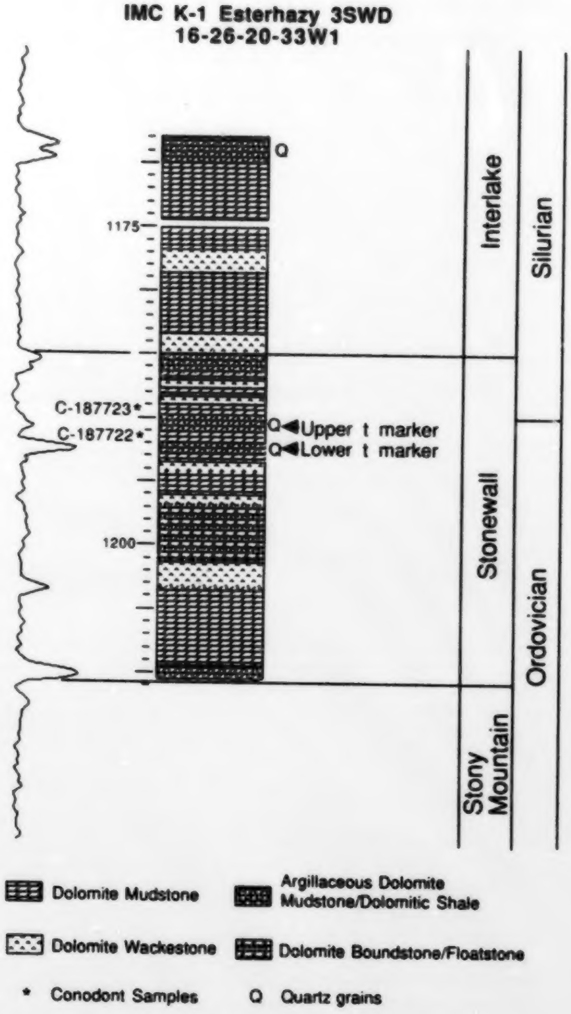


Figure 1 - Gamma ray log and lithology of cored strata encompassing the Ordovician-Silurian boundary in IMC K-1 Esterhazy 3SWD 16-26-20-33W1. Ordovician conodonts were identified in an interval from 1191.38 m to 1191.81 m (GSC loc. C-187722), and Silurian conodonts in a sample taken from 1188.82 m to 1189.47 m (GSC loc. C-187723).

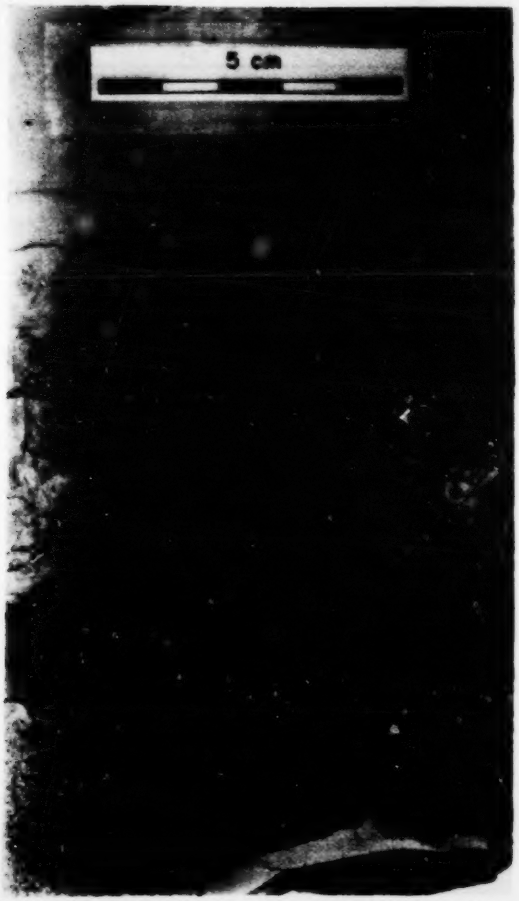


Figure 2 - Contact of dark upper t marker bed with overlying lighter-coloured unit, upper Stonewall Formation. Quartz grains (Q) in the upper t marker bed are concentrated in a lensoid bed at the bottom of the photograph and in disrupted laminae above. The origin of the brecciated texture of the upper part of the marker bed is uncertain. It may be related to solution collapse and/or chemical compaction.

2. Previous work

The precise position of the Ordovician-Silurian boundary within the Stonewall-Interlake sequence has been difficult to establish on the basis of macrofossils due to a paucity of diagnostic fauna. Baillie (1951, p.52) suggested that the Stonewall Formation outcrops in Manitoba "may be transitionary in nature between Richmondian and Alexandrian time". Stearn (1953, 1956), however, concluded that these same strata were entirely Ordovician and that Silurian strata were restricted to the Interlake. His comments were based on outcrops in southern Manitoba that do not include the upper beds found in the subsurface Stonewall Formation. Brindle (1960, p.19) stated that in the subsurface in Saskatchewan, "the Ordovician-Silurian boundary may coincide with the prominent t marker within the Stonewall Formation", based on "rather slender faunal evidence" in a core from Imperial Herald 1-31-1-20W2. He identified Upper Ordovician fossils *Halysites (Catenipora) gracilis* Hall and ?*Opikina stonewallensis* Stearn below the t marker (lower t horizon of Kendall, 1976) and Silurian fossils *Favosites* cf. *favosus* Goldfuss and *Syringopora* sp. in the fossiliferous wackestone unit between the lower and upper t markers in this well.

3. New data

Recent work by the Geological Survey of Canada, in conjunction with the author's ongoing study of Silurian strata in Saskatchewan and her work on the Geological Atlas of the Western Canada Sedimentary Basin, has provided new paleontological data from core of Stonewall and lowermost Interlake strata in the IMC K-1 Esterhazy 3SWD 16-26-20-33W1 well, drilled in 1986 (Figure 1). B.S. Norford examined this core, studied macrofossils, and collected seventeen samples for conodont analyses.

Preliminary conodont data indicate that the Ordovician-Silurian boundary lies within an interval 0.6 m to 2.5 m above the lower t marker (Nowlan and Aldridge, pers. comm.). Ordovician conodonts are present in the interval from 1191.38 m to 1191.81 m (GSC loc. C-187722, above the lower t marker; Figure 1). Silurian conodonts have been identified from 1188.82 m to 1189.47 m (GSC loc. 187723, above the upper t marker; Figure 1). Thus, based on conodont data, the Ordovician-Silurian boundary appears to coincide with the upper t marker, possibly at the top of the uppermost unit which contains abundant sand grains (Figure 2). Macrofossil evidence in this well also supports the placement of the boundary in the vicinity of the t marker: Lower Silurian corals ("*Neozaphrentis*" *hindi* Stearn) are present at 1188.4 m and Upper Ordovician corals *[Paleofavosites* sp., "*Tryplasma*" cf."*T.*" *gracilis* (Whiteaves)] at 1197.6 to 1198.0 m (Norford, pers. comm.).

Recent paleontological studies by the Geological Survey of Canada of outcrop and cores in Manitoba also indicate that the Ordovician-Silurian boundary occurs above the t marker bed (McCabe, 1988; Bezys, 1991).

4. Conclusions

Paleontological evidence in southeastern Saskatchewan places the Ordovician-Silurian boundary between the base of the lower t marker bed and the top of the upper t marker bed. Preliminary conodont data from IMC K-1 Esterhazy 3SWD 16-26-20-33W1 (Nowlan and Aldridge, pers. comm.) indicate that the boundary is associated with the upper t marker whereas macrofossil evidence from Imperial Herald 1-31-1-20W2 (Brindle, 1960) suggests the boundary is associated with the lower t marker.

5. References

Baillie, A.D. (1951): Silurian geology of the Interlake area, Manitoba; Manit. Dep. Mines Nat. Resour., Mines Br. Publ. 50-1, 82p.

Bezys, R.K. (1991): GS-14 Stratigraphic mapping (NTS 63F, 63K) and core hole program 1991; *in* Report of Activities 1991, Manit. Energy Mines.

Brindle, J.E. (1960): The faunas of the Lower Paleozoic carbonate rocks in the subsurface of Saskatchewan; Sask. Dep. Miner. Resour., Rep. 52, 45p.

Johnson, M.E. and Lescinsky, H.L. (1986): Depositional dynamics of cyclic carbonates from the Interlake Group (Lower Silurian) of the Williston Basin; Palaios, v1, p111-121.

Kendall, A.C. (1976): The Ordovician carbonate succession (Bighorn Group) of southeastern Saskatchewan; Sask. Dep. Miner. Resour., Rep. 180, 185p.

McCabe, H.R. (1988): Preliminary report on Ordovician-Silurian boundary rocks in the Interlake area, Manitoba, Canada; Bull. Br. Mus. Nat. Hist. (Geol), v43, p255-257.

Norford, B.S., Haidl, F.M., Bezys, R.K., Cecile, M.P., McCabe, H.R., and Paterson, D.F. (in press): Middle Ordovician to Lower Devonian strata of the Western Canada Sedimentary Basin; *in* Mossop, G.D. (ed.), Geological Atlas of the Western Canada Sedimentary Basin.

Osadetz, K.G. and Haidl, F.M. (1989): Tippecanoe Sequence: Middle Ordovician to lowest Devonian: vestiges of great epeiric sea; *in* Ricketts, B.D. (ed.), Western Canada Sedimentary Basin: A Case Study, Can. Soc. Petrol. Geol., p121-137.

Porter, J.W. and Fuller, J.G.C.M. (1959): Lower Paleozoic rocks of northern Williston Basin; Am. Assoc. Petrol. Geol., Bull., v43, p124-189.

Stearn, C.W. (1953): Ordovician-Silurian boundary in Manitoba; (abstract). Bull. Geol. Soc. Am., New York, v64, p1477-1478.

__________ (1956): Stratigraphy and paleontology of the Interlake Group and Stonewall Formation of southern Manitoba; Geol. Surv. Can., Mem. 281, 162p.

Preliminary Palynological Investigations of Upper Cretaceous Strata, Southwestern Saskatchewan

C.F. Gilboy and E.M.V. Nambudiri[1]

Gilboy, C.F. and Nambudiri, E.M.V. (1991): Preliminary palynological investigations of Upper Cretaceous strata, southwestern Saskatchewan; *in* Summary of Investigations 1991, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 91-4.

Twelve core samples from four wells penetrating Upper Cretaceous strata in southwestern Saskatchewan were analyzed for their palynomorph contents. The main aim of the investigation is to learn whether palynological assemblages obtained from core are helpful in establishing detailed age relationships of source rocks which generally lack identifiable megafossils. In addition, the use of palynomorphs in supplementing sedimentological information about depositional environments is assessed.

1. Results

Cored intersections from two of the wells are of Second White Speckled Shale, one (06-25T-14-30W3) in the basal sand unit and an associated bentonite (B bentonite, Gilboy, 1988a, b), the other (11-28-02-28W3) in the uppermost sandy unit and a distinctive regionally recognized bentonite (C bentonite, Gilboy, 1988a,b).

a) NCO Hatton 06-25T-14-30W3, Four Samples

Sample 1, at 667.18 m (sample 1 of Nambudiri, 1991): dark grey to black, finely laminated shales with rare, siltstone laminae (less than 1 mm thick); very little bioturbation; sporadic fish-scales and spines, and 1 to 2 cm long casts of unbroken pelecypod valves. The palynomorph assemblage comprises: *Cicatricosisporites* sp., *Cicatricosisporites augustus* Singh, *Cyathidites minor* Couper, *Cyathidites australis* Couper, *Gleicheniidites senonicus* Ross, *Lycopodiumsporites marginatus* Singh, *Appendicisporites bilateralis* Singh, *Trilobosporites apiverrucatus* Couper, *Biretisporites potoniaei* Delcourt and Sprumont, *Cycadopites* sp., *Podocarpidites granulatus* Singh, *Alisporites bilateralis* Rouse, *Laricoidites magnus* (Potonie) Potonie, Thomson and Thiegart, *Cerebropollenites mesozoicus* (Couper) Nilsson , *Abiespollenites* sp., *Taxodiaceaepollenites hiatus* (Potonie) Kremp, *Clavatipollenites hughesii* Couper, *Liliacidites inequalis* Singh, Tricolporate pollen, *Tricolpites sagax* Norris, and abundant dinoflagellates and acritarchs. The presence of *Taxodiaceaepollenites hiatus, Cicatricosisporites* sp., *Gleicheniidites senonicus, Alisporites bilateralis* and angiosperm pollen such as *Tricolpites sagax* suggests possible affinities of this assemblage with the early Tricolpate suite of Norris *et al.* (1975), the upper part of which has been assigned a Cenomanian age. The abundance of black wood indicates offshore deposition or oxidizing environments.

Sample 2, at 665.07 m (sample 3 of Nambudiri, 1991): parallel laminated fine sandstones and shales, commonly graded in 1 to 3 cm thick units some of which are moderately bioturbated, others are non-bioturbated; immediately overlying these is a 15 cm thick unit of black shale, possibly representing a flooding surface. The palynomorph assemblage is: *Cyathidites minor* Couper, *Trilobosporites tribotrys* Dettmann, *Gleicheniidites senonicus* Ross, *Deltoidospora hallii* Miner, *Taxodiaceaepollenites hiatus* (Potonie) Kremp, *Laricoidites magnus* (Potonie) Potonie Thomson and Thiegart, *Vitreisporites pallidus* (Reissinger) Nilsson, *Podocarpidites granulatus* Singh, *Cedripites canadensis* Pocock, *Tricolpites* sp., *Tricolpites parvus* Stanley, *Liliacidites magnus* Singh, *Betulaceoipollenites* sp., *Cupuliferoidaepollenites minutus* (Brenner) Singh, *Clavatipollenites minutus* Brenner, *Veryachium* sp., and abundant dinoflagellates and hystrichosphaerids. The continued appearance of *Taxodiaceaepollenites hiatus* and *Gleicheniidites senonicus* along with the angiosperm pollen *Cupuliferoidaepollenites minutus, Liliacidites* sp. and *Tricolpites* sp. again suggests affinities of this assemblage with the early Tricolpate suite of Norris *et al.* (1975). However, major differences exist, and the lack of diagnostic palynomorphs make age determination for this sample imprecise. Continentally derived bisaccate conifer pollen is abundant, and may have been transported a long distance. Both black and brown woods are present.

Sample 3, at 662.62 m (sample 5 of Nambudiri, 1991): finely laminated black shale with rare siltstone laminae and lenses, very weakly bioturbated (several subhorizontal, sand-filled tubes are present). This unit separates intensely *Terebellina*-bioturbated, medium grey sandstone below from overlying, weakly bioturbated, greenish, locally ripple cross-bedded fine sandstones and interbedded shales. No palynomorphs were recovered.

Sample 4, at 660.92 m (sample 4 of Nambudiri, 1991): very weakly bioturbated to non-bioturbated, greenish, fine, thinly laminated (laminae up to about 1 mm thick), ripple cross-bedded sandstones with thin shale partings; beds are mostly 1 to 2 cm thick, and many are normally graded with coarser-grained basal parts cemented by carbonate. No palynomorphs were recovered.

---

(1) Department of Geology, University of Regina, Regina, Saskatchewan

Sample 5, at 659.18 m (sample 2 of Nambudiri, 1991): dark grey to black, finely laminated shales with rare, siltstone laminae less than 0.5 mm; small flakes of mica (including biotite) are present, as are sporadic fish remains. No palynomorphs were recovered.

b) United Bata Resources Noel Senate 11-28-02-28W3, Three Samples

Sample 6, at 819.06 m (sample 8 of Nambudiri, 1991): interlayered spotted and non-spotted, dark grey shales with rare lenticles of fine sandstone up to 3 mm thick; weakly bioturbated (subhorizontal, sand-filled burrows); contains biotite flakes, abundant plant remains, and fish-scales. The palynomorph assemblage includes *Appendicisporites* sp., *Deltoidospora hallii* Miner, *Concavissimisporites punctatus* (Delcourt and Sprumont) Singh, *Taxodiaceaepollenites hiatus* (Potonie) Kremp, *Podocarpidites* sp., *Abiespollenites* sp., and numerous dinoflagellates and organic-walled microforaminifera. The lack of age-diagnostic palynomorphs and the limited variety of species preclude correlation with other assemblages. The presence of microforaminifera and dinoflagellates suggests marine depositional environments.

Sample 7, at 804.73 m (sample 6 of Nambudiri, 1991): moderately speckled, dark grey shales containing fish-scales and -spines as well as a layer of parallel *Baculites* shells (mostly at least 9 cm in length) and sporadic unbroken pelecypod valves up to 2 cm across. The base of the 64 cm thick C bentonite (Gilboy, 1988a,b) is present 82 cm above the sample location, which is situated 5 cm below a scour surface where low-angle cross-bedded sand laminae less than 1 mm thick are eroded and overlain by shale. No palynomorphs were recovered.

Sample 8, at 800.77 m (sample 10 of Nambudiri, 1991): richly speckled medium to dark grey shales with fish debris, including vertebrae; contain interlaminated carbonate-rich siltstone layers, generally less than 1 mm thick; the sample location is 2.4 m above the top of the 64 cm thick C bentonite. The palynomorph assemblage is made up of *Deltoidospora hallii* Miner, *Cyathidites minor* Couper, *Concavissimisporites* sp., *Concavissimisporites variverrucatus*. (Couper) Brenner, *Trilobosporites apiverrucatus* Couper, *Cycadopites* sp., *Podocarpidites radiatus* Brenner, *Podocarpidites granulatus* Singh, *Taxodiaceaepollenites hiatus* (Potonie) Kremp, *Laricoidites* sp., *Abiespollenites* sp., *Classopolles classoides* Pflug emend. Pocock and Jansonius, *Alisporites* sp., *Tricolpites sagax* Norris, *Tricolpites reticulatus* Cookson ex Couper emend. Potonie, *Retitricolpites vulgaris* Pierce, *Clavatipollenites* sp., *Liliacidites* sp. (abundant), *Odontochitina* sp., numerous dinoflagellates and very abundant hystrichosphaerids. Although no precise age can be assigned to this assemblage, it can be compared with the early Tricolpate suite of Norris *et al.* (1975). Species of *Odontochitina* are commonly Albian in age, but Norris *et al.* have found specimens in an assemblage of probable Cenomanian age. Black woods are abundant in this sample, brown woods rare. Bisaccate grains indicate long distance transport.

c) CDN Roxy AEC Cypress 07-22-09-28W3, Two Samples

This well was cored in the Medicine Hat Sandstone.

Sample 9, at 639.03 m (sample 12 of Nambudiri, 1991): highly speckled, medium grey, laminated fine sandstone and shale, sand laminae up to 2 mm thick. The palynomorph assemblage is: *Gleicheniidites senonicus* Ross, *Alisporites* sp., *Cycadopites* sp., *Liliacidites magnus* Singh, and *Tricolpites microreticulatus* Belsky, Boltenhagen and Potonie. The lack of index palynomorph species preclude age assignments and comparison with other assemblages.

Sample 10, at 634.73 m (sample 11 of Nambudiri, 1991): highly speckled, medium grey shale with fish-scales up to about 3 mm across along with other fish debris including bones (vertebrae) and spines up to 5 mm long. The palynomorph assemblage includes *Gleicheniidites senonicus* Ross, *Appendicisporites* sp., *Cyathidites minor* Couper, *Taxodiaceaepollenites hiatus* (Potonie) Kremp, *Trudopollis* sp., *Retitricolpites vulgaris* Pierce, *Proteacidites retusus* Andersen, *Tubuliferoidites* sp., *Clavatipollenites hughesii* Couper, *Tricolpites reticulatus* Cookson ex Couper emend. Potonie, *Cercidiphyllites brevicolpites* Mtchedlishvili, *Betulaceaoipollenites* sp., *Dinogymnyum acuminatum* Evitt *et al.*, *Varyachium* sp., and an abundance of hystrichosphaeridia and other dinoflagellates.

This palnomorph assemblage is typical of the early Triporate suite of Norris *et al.* (1975), to which the basal part of the Lea Park Formation in Alberta belongs. Several redeposited spores were identified.

d) City of Medicine Hat Hatton 08-15-16-28W3, Two Samples

This well was cored in the Milk River Formation.

Sample 11, at 500.3 m (sample 7 of Nambudiri, 1991): medium grey, poorly sorted, generally fine-grained, impure sandstone containing about 50 percent clay minerals as well as flakes of biotite and white mica; abundant small fragments of carbonized wood impart a salt-and-pepper appearance to the rock, which has a generally massive texture suggestive of intense bioturbation; isolated shale lenses are present. The palynomorph assemblage is made up of *Trilobosporites trioreticulosus* Cookson and Dettmann, *Gleicheniidites senonicus* Ross, *Deltoidospora hallii* Miner, *Cyathidites minor* Couper, *Lycopodiumsporites marginatus* Singh, *Laevigatosporites ovatus* Wilson and Webster, *Concavissimisporites* sp., *Taxodiceaepollenites hiatus* Brenner, *Laricoidites* sp., *Podocarpidites radiatus* Brenner, *Tricolpites microreticulatus* Belsky, Boltenhagen and Potonie and *Clavatipollenites hughesii* Couper. It is poorly preserved, and the lack of characteristic index species prevents age and environmental interpretations.

Sample 12, at 436.61 m (sample 9 of Nambudiri, 1991): Medium grey, fine, impure sandstone (up to 70 percent clay content) with numerous shell fragments showing an iridescent sheen. The palynomorph assemblage comprises *Lycopodiumsporites* sp., *Deltoidospora hallii* Miner, *Gleicheniidites senonicus* Ross, *Gleicheniidites bolchovitinae* Doring, *Lophotriletes babase* (Brenner) Singh, *Dictyophyllidites* sp., *Ciratriradites teter* Norris, *Cicatricosisporites hallii* Delcourt and Sprumont, *Acanthotriletes* sp., *Cyathidites minor* Couper, *Biretisporites potoniaei* Delcourt and Sprumont, *Trilobosporites* sp., *Osmundacidites* sp., *Taxodiaceaepollenites hiatus* (Potonie) Kremp, *Podocarpidites granulatus* Singh, *Alisporites bilateralis* Rouse, *Abiespollenites* sp., *Cibotiidites arlii* Srivastava, *Triatriopollenites costatus* Norton, *Aquilapollenites calvus* Tschudy and Leolpold, *Aquilapollenites quadrilobus* Rouse, and abundant dinoflagellates and hystrichosphaerids. It is comparable with the early loranthaceous suite of Norris *et al.* (1975) in Alberta, which contains pteridophyte spores such as *Cicatricosisporites hallii*, *Cyathidites minor*, *Deltoidospora hallii* and *Gleicheniidites senonicus*, and conifer pollen (*Abiespollenites* sp. and *Taxodiaceaepollenites hiatus)*. The occurrence of species of *Aquilapollenites, especially A.calvus* and *A. quadrilobus*, is particularly significant in correlating this sample with the early loranthaceous suite. Norris *et al.* (1975) proposed this suite to be characteristic of the middle part of the Lea Park Formation. They suggested a Santonian to Campanian age for this formation, but recent dinoflagellate studies by Harker *et al.* (1990) indicate a Campanian age. This sample contains much opaque to dark brown and brown wood, as well as an abundance of spores and pollen (including bisaccate conifer pollen), dinoflagellates and acritarchs. The black wood indicates offshore environments (Cross *et al.*, 1966; Muller, 1959) or oxidizing conditions (Pocock *et al.*, 1987-1988). Melia (1982) suggested that terrestrial palynomorphs (pollen and spores) decrease in abundance away from continental margins towards shelf and slope, where sediments contain an admixture of pollen, spores, dinoflagellates and acritarchs.

2. Conclusions

Of the samples used in this study, palynomorphs were not recovered from four, three yielded palynomorph assemblages too mediocre to allow for age or environmental interpretations, and five provided data of some use in defining the age and/or the depositional environment of the host rocks. Cenomanian ages indicated for Second White Speckled Shale samples (1, 2 and 8) are in agreement with expected results (Gilboy, 1987), as is a Campanian age for the uppermost Milk River Formation (12). Both the basal Lea Park Formation and the uppermost Medicine Hat Sandstone (10) are generally considered to be of Santonian age. Palynomorph assemblages and preserved wood remains indicated that the host rocks were originally deposited in outer shelf environments. Assemblages in southwestern Saskatchewan are closely comparable with those at similar stratigraphic levels in Alberta.

Reconnaissance-level palynomorph studies of Upper Cretaceous clastic sedimentary rocks from southwestern Saskatchewan are of broad use in defining age relationships and depositional environments.

3. References

Cross, A.T., Thompson, G.G. and Zaitzeff, J.B. (1966): Source and distribution of palynomorphs in bottom sediments, southern part of the Gulf of California; Marine Geol., v4, p467-524.

Gilboy, C.F. (1987): Aspects of the regional geological framework of low-permeability shallow gas reservoirs in Upper Cretaceous strata (Colorado and Montana Groups), southwestern Saskatchewan; *in* Summary of Investigations 1987, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 87-4, p199-213.

___________ (1988a): Geology and natural gas production of the Upper Cretaceous Second White-speckled Shale, southwestern Saskatchewan; *in* Summary of Investigations 1988, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 88-4, p183-195.

___________ (1988b): South-north stratigraphic cross-section through Upper Cretaceous Second White-speckled Shale in southwest Saskatchewan; *with* Summary of Investigations 1988, Saskatchewan Geological Survey, Sask. Energy Mines, Misc. Rep. 88-4.

Harker, S.D., Sargeant, W.A.S. and Caldwell, W.E.G. (1990): Late Cretaceous (Campanian) organic-walled microplankton from the Interior Plains of Canada, Wyoming and Texas: Biostratigraphy, palaeontology and palaeoenvironmental interpretation; Palaeontographica B, v219, p1-243.

Melia, M.B. (1982): The distribution and relationship between palynomorphs in aerosols and deep-sea sediments off the coast of northwest Africa; Palynol., v6, abstr., p285.

Muller, J. (1959): Palynology of Recent Orinoco delta and shelf sediments; Micropaleont., v5, p1-32.

Nambudiri, E.M.V. (1991): Preliminary investigation of the palynology of the Cretaceous strata of southwestern Saskatchewan; U. Regina, Energy Resear. Unit Contrib. No. 248, 11p.

Norris G., Jarzen, D.M. and Awai-Thorne, B.V. (1975): Evolution of the Cretaceous terrestrial palynoflora in western Canada; *in* Caldwell, W.G.E. (ed.), The Cretaceous System in the Western Interior of North America; Geol. Assoc. Can., Spec. Pap. 13, p333-364.

Pocock, S.A.J., Vasanthy, G. and Venkatacha-la, B.S. (1987-1988): Introduction to the study of particulate organic materials and ecological perspectives; J. Palynol., v23-24, p167-188.

F. Appendices

Titles and Abstracts of Recent Saskatchewan Energy and Mines Publications and Papers by Staff of the Saskatchewan Geological Survey

---

1. Reports

Kreis, L.K. (1991): Stratigraphy of the Jurassic System in the Wapella-Moosomin area, southeastern Saskatchewan; Sask. Energy Mines, Rep. 217, 90p.

Scott, B.P. (1991): Metallogenic Map Series: Reindeer Lake South NTS area 64D; Sask. Energy Mines, Rep. 253, 18p, (accompanied by 1:250 000 scale map).

2. Miscellaneous Reports

Mineral Resources of Saskatchewan; Sask. Energy Mines, Misc. Rep. 91-5, 76p.

Bennett, R.W. (1991): Assessment Work Catalogue, Saskatchewan; Sask. Energy Mines, Misc. Rep. 91-6, 96p.

3. Open Files

Coombe Geoconsultants Ltd. (1991): Base Metals in Saskatchewan; Sask. Energy Mines, Open File Rep. 91-1, 218p.

Pearse, G.H.K. (1991): Building Stone Potential in the Creighton-Pelican Narrows Region of Saskatchewan; Sask. Energy Mines, Open File Rep. 91-3, 32p.

Roberts, R.G. (1990): Structural Controls of the Box and Athona Deposits, Beaverlodge Area, Saskatchewan; Sask. Energy Mines, Open File Rep. 90-2, 17p.

4. Miscellaneous Publications

Mineral Resource Map of Saskatchewan (1990): Sask. Energy Mines, 1990 Edition (in large (71 x 101 cm) and small (28 x 38 cm) formats).

5. External Papers

Ashton, K.E. and Leclair, A.D. (1991): Foliate: a useful term to complement the textural classification of foliated metamorphic rocks: Reply; Can. J. Earth Sci., v28, p829-831.

Delaney, G.D. (1991a): Gold in the Glennie Domain, Lower Proterozoic Trans-Hudson Orogen, northern Saskatchewan; The Geological Society of CIM, First Annual Field Conference, Sept. 1991, Saskatoon, Prog. Abstr., p11-12.

When it is placed in production in November 1991, the Seabee deposit will become Saskatchewan's fifth and potentially largest gold mine. It will also be the first gold mine in the underrated and commonly overlooked Glennie Domain which encompasses over 40 gold occurrences, many discovered in recent years. All known gold showings are in or adjacent to greenstone belts. By far the greatest number, including the Seabee deposit, have been found in the extensive Pine Lake Greenstone Belt. For all greenstone belts which have been investigated the volcanics are primarily of tholeiitic composition and were formed in juvenile island arc settings. Host rocks comprise volcanics, plutons including both subvolcanic as well as younger post-volcanic suites, volcaniclastics, sediments and migmatites. Metamorphic grade is upper greenschist to amphibolite facies.

Most gold showings are structurally controlled (i.e. Seabee, Georges Lake and Manawan); a few are intrusive hosted (i.e. Kalix, and Jena around the Brownell Lake Pluton) and there is one example of a sediment-hosted occurrence (Pine Lake Joint Venture). Gold also occurs in association with volcanogenic massive sulphides. Structurally controlled mineralization typically occurs in secondary or higher order structures, proximal to major discordances, where they transect zones of lithologic heterogeneity. Gold is found within and adjacent to quartz veins; many of these are multistage although not all stages are mineralized. The element association of the gold-quartz veins is variable. There do, however, appear to be some locally distinctive associations. For example, tourmaline is found in most veins within or adjacent to the Laonil Lake intrusive complex; arsenopyrite is characteristic of showings (both structural and plutonic hosted) in the Brownell-Wapawekka Lakes Greenstone Belt. Timing of mineralization has varied. The Glennie Domain has a high potential for the discovery of other significant gold deposits.

Delaney, G.D. (1991b): Footwall Fragmentals: A Key to understanding the evolution of the feeder system to the Sullivan orebody; The Geological Society of CIM, First Annual Field Conference, Sept. 1991, Saskatoon, Prog. Abstr., p37-38.

The Sullivan deposit lies near the top of the Middle Proterozoic, Lower Aldridge Formation, a sequence of predominantly fine-grained, argillaceous siliciclastic sediments which accumulated in an abyssal setting. The 160 million tonne Sullivan massive sulphide deposit is approximately 2000 m long in a north-south direction by 1600 m wide, and has the shape of a convex lens that dips moderately east. It can be subdivided into two parts: a western, massive part which is up to 100 m thick and, separated by a transition zone, an eastern bedded part which consists of five bedded sulphide intervals intercalated in sedimentary rocks. A large volume of fragmental rock occurs in the footwall to the Sullivan deposit and defines the first stage in the evolution of the feeder system; a minor amount of these rocks is found in the hanging wall.

There are two textural varieties of footwall fragmental. Pebble fragmental, the most abundant type, comprises subangular to subrounded, granule- to predominantly pebble-sized fragments, forming a condensed to disrupted framework. The second type, chaotic breccia, consists of angular, granular- to boulder-size fragments forming a generally intact to locally disrupted framework. Underneath the southern half of the orebody, pebble fragmental occurs in several discrete bodies most of which crosscut stratigraphy. Fragmental dykes range in height and width from tens of centimetres to tens of metres, pinch and swell in thickness and locally bifurcate. In some dykes, clasts are characterized by wispy terminations. The lithologic spectrum of clasts matches that of the encompassing Lower Aldridge sediments although many of these clasts have been subjected to subsequent tourmalinization, albitization or chloritization. Locally near the margins of the fragmental bodies there are angular boulder size fragments which are of the same composition as the surrounding wall rocks. In the bedded sequence, below some of the cross-cutting fragmentals, there are commonly micro-faults and sand or silt dykelets; in rare cases stratiform apophyses occur at the top of the cross-cutting fragmentals.

Beneath the northern half of the orebody, to the west of the transition zone, the pebble fragmental is a more or less continuous body characterized by an irregular top and base as well as by variable thickness. The basal contact is slightly unconformable to distinctly cross-cutting and in the footwall there are commonly cross-cutting fragmental dykes. Also within this basal zone, some areas of angular fragments up to a few metres wide contain fragments with bedding in the same attitude as the adjacent undisturbed sediments. Elsewhere bedding has been rotated. The top of this part of the pebble fragmental is also highly irregular. Locally there is a conformable contact with overlying sediments of the Lower Aldridge Formation with up to 15 m of sediments occurring between the top of the fragmental and the sulphide footwall. More commonly, however, an irregular hummocky discordance characterizes the top of the fragmental and at several localities the sulphide footwall contacts the fragmental. Textural variations are illustrated by a framework ranging from condensed to disrupted and clasts from granule to pebble size. Locally, at the base, near the sides and in zones where chaotic breccia forms an admixture with the pebble fragmental, there are larger fragments and in places particularly near the contacts, there is a well developed clast fabric.

East of the transition zone, underneath the northern part of bedded ore, the pebble fragmental is a more regular, generally stratiform body characterized by an intact framework. Although most clasts are of fine-grained sedimentary rocks of the Lower Aldridge Formation, there are also rare isolated tourmalinite clasts. This stratiform fragmental, which is up to 30 m thick near the transition zone, gradually thins to the northeast. The basal contact of the fragmental sheet varies from conformable to slightly disconformable. Immediately east of the transition zone, there is up to 15 m of bedded sediment between the top of the fragmental sheet and the sulphide footwall; the thickness of this bedded interval gradually decreases to the northeast until sulphide rests directly on fragmental. Along its southern boundary the thickness of the stratiform pebble fragmental decreases abruptly; to the north relationships are masked by subsequent faulting.

A model for the early evolution of the Sullivan deposit is presented. The Lower Aldridge Formation, a sequence of predominantly fine-grained sediments, was deposited in a basinal setting. Elevated pore pressure developed in the sediment pile; contributing factors may have included rapid rate of sedimentation, an increase in thermal gradient and methane generation by anaerobic bacteria. Abrupt release of pore pressure formed cross-cutting fragmental dykes, some of which vented at the sediment water interface forming local slump deposits. Beneath what would become the western part of the orebody, growth and amalgamation of fragmental dykes resulted in a large mound of variably fragmented sediments which subsequently slumped into a northeast tending trough. During this phase an initial restricted pulse of tourmalinization occurred. Slumping was followed by subsidence of the core of the western fragmental complex and sedimentation continued. Rejuvenation of the core zone of the fragmental complex resulted in the development of new zones of fragmentation reaching the sediment water interface concurrently with the influx of tourmaline-rich brines. This was followed by the passage of metal rich brines which formed the orebody at or very close to the sea floor.

Delaney, G.D., Czornobay, B. and Earle, S. (1991): Utilization of image processed aeromagnetic and Landsat thematic image data for the improved understanding of the lithologic and structural characteristics of the Glennie Domain; The Geological Society of CIM, First Annual Field Conference, Sept. 1991, Saskatoon, Prog. Abstr., p11.

The Saskatchewan Geological Survey has completed a pilot study which investigated the use of image processed aeromagnetic data and filtered Landsat thematic mapper data for enhanced understanding of the lithologic and structural characteristics of the Glennie Domain. This survey also helped identify structural and lithologic features which are important from a mineral exploration perspective. Enhancement of aeromagnetic data, including the creation of multi-colour magnetic anomaly maps and shadow images was achieved by utilizing the software package RTI-CAD of GEOPAK SYSTEMS of Toronto. Landsat thematic image data was processed through digital lineament detection filters in four directions with the ARIES-II software of DIPLIX Technologies of Ottawa. RTI-CAD was also used to create multi-layered images that integrated enhanced aeromagnetic, Landsat lineament and geological maps.

Several data sets were utilized. Aeromagnetic data from federal-provincial surveys flown at a line spacing of approximately 400 m was regridded to 200 m and then processed for the whole Glennie Domain. From the resulting magnetic anomalies and shadow maps it was possible to define major scale structural features and the morphology of many lithologic entities. For example, basic to ultrabasic intrusives in the central part of the domain, most of which contain structural-hosted gold showings, are defined by magnetic highs. These intrusives include the Laonil Lake intrusive complex, the Eisler Lake intrusion, the Leland Lake metagabbro and the Robinson Lake amphibolite. Detailed proprietary aeromagnetic surveys, flown at line spacings of 100 m to 200 m were processed for two areas: around Brownell Lake and between Laonil Lake and the junction of the Churchill and Reindeer Rivers. The resolution of features from the processed data of these surveys was greater and it was possible to delineate higher order structures and subtle lithologic features not apparent from the federal-provincial surveys flown at a wider flight line spacing. As an example, the major east-west trending Hartley Shear zone, which together with subsidiary faults has played an important role in localizing gold mineralization in the Brownell Lake area, is defined by a narrow, linear magnetic high.

Additional understanding of the character and distribution of high strain zones was obtained by integrating enhanced magnetic data with lineament maps derived from Landsat thematic mapper data and geologic maps. Northeast- to southeast-trending high strain zones of various magnitudes and which apparently formed late in the thermotectonic history of the domain, played a key role in localizing gold mineralization.

Froese, E. and Ashton, K.E. (1991): Alteration zones in the Kisseynew Gneisses of Saskatchewan and Manitoba; The Geological Society of CIM, First Annual Field Conference, Sept. 1991, Saskatoon, Prog. Abstr., p25-26.

The Kisseynew gneisses represent metamorphosed equivalents of lower-grade rocks present in the Flin Flon volcanic belt. In recent work, the Kisseynew gneisses have been mapped using the stratigraphic subdivisions established in the Flin Flon belt. A noteworthy feature is the presence of felsic volcanic rocks, previously regarded as metasediments. There are numerous occurrences of altered rocks; the precursors are thought to be rocks hydrothermally altered prior to metamorphism. Four types of alteration can be distinguished: 1. carbonatization (calc-silicate rocks); 2. moderate Fe-Mg metasomatism (cummingtonite-bearing rocks); 3. intense Fe-Mg metasomatism (cordierite-anthophyllite rocks); and 4. aluminous alteration (sillimanite-bearing rocks).

Sulphide deposits in carbonatized rocks commonly consist mainly of pyrite and pyrrhotite. Some rocks indicating Fe-Mg metasomatism and aluminous alteration are associated with base metal sulphides. Any possible zonation of alteration types has been obscured by deformation. The recognition of a greater proportion of volcanic precursors and the presence of altered rocks enhance the exploration potential of a substantial area in Saskatchewan and Manitoba.

Gent, M.R. (1991a): Diamonds in Saskatchewan; Alta. Geol. Surv. and B.C. Energy, Mines Petrol. Resour., 27th Forum on the Geology of Industrial Minerals, May 1991, Banff, Prog. Abstr.

Gent, M.R. (1991b): Diamond Exploration in Saskatchewan; CIMM 93rd Annual General Meeting, May 1991, Vancouver, Prog. Abstr.

The diamond industry has an attractive long term past and projected market stability. The exploration and mining costs for diamonds tend to be low. The industry has a high success factor in comparison to other mineral exploration.

The presence in Saskatchewan of eight kimberlite pipes of probable late Albian age has been reported. An abundance of aeromagnetic anomalies possibly attributable to kimberlite pipes have been identified and it is anticipated that industry has acquired reasonable proof of the presence of twenty to forty or more kimberlite pipes in Saskatchewan. Historical reports of diamonds, the discovery of kimberlite pipes and these aeromagnetic anomalies indicate the Phanerozoic Basin and some parts of the exposed Shield in Saskatchewan have an excellent potential for the occurrence of economically exploitable diamond deposits. The presence of some igneous minerals found in some salt solution breccias also indicates the probable occurrence of extensive diatreme activity during the Devonian.

A Laramide Orogeny-induced crustal hinging or arching of the eastern margin of the Phanerozoic Basin and probably dextral wrenching-induced arching and faulting within the basin are proposed as major factors in the emplacement of Cretaceous or younger kimberlites. Zones of crustal weakness between subcratonic Archean blocks and internal faults are also thought to be significant controlling factors in the location of kimberlites in Saskatchewan.

Mapped aeromagnetic anomalies interpreted as being possibly due to kimberlite pipes generally tend to be concentrated in areas underlain by a basement of or with components of Archean age. Within Saskatchewan, these anomalies occur within the probable Trans-Hudson Orogen zone, the Rae-Hearn Craton margin, and the northeast flank of the Elkpoint Basin. Recognized areas of potential for kimberlite emplacement are the Phanerozoic basin margin, the western limit of the Superior Province adjoining the Churchill Province, and a swath approximately 150 km wide along the NE-SW trending Shaunavon linear and other areas of inner basin arching.

Harper, C.T. (1991): The Deception Lake lead-zinc discovery; A base metal enriched iron formation; The Geological Society of CIM, First Annual Field Conference, Sept. 1991, Saskatoon, Prog. Abstr. p10.

Lead-zinc (± silver, gold) mineralization was discovered in 1989 near Northwest Bay of Deception Lake, 200 km north-northeast of La Ronge, Saskatchewan. The discovery was made in a little explored part of the Rottenstone Domain which comprises trondhjemitic granitoid and migmatitic metasedimentary gneisses and the extensive Wathaman Batholith. Northeast Bay lies in a transitional zone between granitoids to the northwest and metasediments to the southeast. Paleosomal remnants rarely compose more than twenty-five percent of the rocks, and include: psammitic to pelitic and amphibolitic gneisses and cal-silicate rocks. The most economically significant rock type is a calcareous, silicate-sulphide facies iron formation assemblage which hosts the lead-zinc-silver-gold mineralization. These rocks occupy an elongate, northeast-trending, overturned, moderately northwest-dipping, and shallow to moderately northeast-plunging, synformal D_2 fold. Syntectonic pegmatitic trondhjemite sheets are generally parallel to the axial surface of the regional fold.

The iron formation, which is found throughout the area, typically consists of diopside-quartz-plagioclase ± garnet ± graphite with an average of 10% finely disseminated pyrrhotite, but locally contains pods and lenses of massive pyrrhotite. Gahnite (zinc spinel), pyrite and chalcopyrite are rare. Thickness and composition of the iron formation is very uniform and can be traced by mapping and geophysics for many kilometres. The lead-zinc mineralization is associated with minor interlayered biotite and amphibolite schists, biotitic and sulphide-bearing diopsidites (calc-silicates) and impure marbles, which have a more restricted distribution. Pegmatitic trondhjemite, which intrudes the iron formation, is commonly tinted green and contains coarse pyrite, galena and sphalerite locally.

Galena and sphalerite occur as disseminations but more typically occur as millimetre to centimetre thick layers alternating with siliceous and calc-silicate layers. Sphalerite concentrations are greatest in the diopsidite rocks, whereas the highest silver and gold values are associated with the galena-rich sections.

The stratiform and layered character of the mineralization and its association with a regionally extensive iron-rich sedimentary assemblage (i.e. an iron formation) indicates that it formed in a sedex type environment. The skarn-type mineral assemblage is characteristic of calcareous iron silicate facies iron formation assemblages that developed during regional metamorphism, not by contact metamorphism. The Deception Lake iron formation and base metal mineralization is not unusual. Similar rocks and occurrences can be found elsewhere in the Rottenstone Domain (Zn-Cu Brown claim, Paskwachi Bay, Reindeer Lake), in the Crew Lake Belt (Zn-Pb-Ag Borys Lake prospect) and MacLean Lake Belt (Zn-Cu McKenzie deposit, Brabant Lake) of the La Ronge Domain and probably the Kisseynew Domain. This group of stratiform base metal deposits associated with iron-rich sediments has many world class deposits (e.g. Broken Hill, Sullivan, Aggenays and Gamsberg). Thus, a potential exists for finding economic base metal deposits of this type.

Kreis, L.K. (1991): Depositional history of the Jurassic System and hydrocarbon accumulations in the Wapella-Moosomin area, southeastern Saskatchewan; in Christopher, J.C. and Haidl, F.M. (eds.), Sixth International Williston Basin Symposium, Sask. Geol. Soc., Spec. Publ. No. 11, p165-178.

The Triassic/Jurassic section of southeastern Saskatchewan is an unconformity bounded sequence between the Mississippian subcrop and the Cretaceous Mannville Formation. Four lithostratigraphic units are recognized in the Jurassic of southeastern Saskatchewan: the Watrous, Gravelbourg, Red Jacket and Rierdon formations. The Red Jacket is the facies equivalent to strata between the lower member of the Gravelbourg and the Rierdon Formation.

The Triassic/Jurassic(?) red beds of the lower member of the Watrous Formation represent the earliest post-Mississippian deposits in the study area. They appear to have been laid down in a restricted marine environment but with repeated exposure, oxidization, and reworking. This phase was followed in the Bajocian by less restricted marine deposition, represented by anhydrites, dolomites and dolomitic mudstones of the upper member of the Watrous Formation and dolomitic limestones (unit A) of the lower member of the Gravelbourg Formation. Interbedded clastics and carbonates (Unit B) of the lower Gravelbourg Formation were deposited in a near-normal marine environment. In contrast, clastics and subordinate carbonates of the succeeding Red Jacket Formation were deposited during the Bajocian and Bathonian in depositional environments ranging from freshwater to brackish-water to open marine. This variation was probably caused by minor yet frequent fluctuations in sea level, causing broad shifts in the depositional environments. The unstable period of Red Jacket deposition was followed by a major marine transgression and shales of the Rierdon Formation were laid down in the study area during the Bathonian to Callovian.

Most Jurassic hydrocarbon production in southeast Saskatchewan is from the Rocanville and Moosomin members of the Red Jacket Formation in the Wapella-Moosomin area. Oil shows are also in the lower member of the Gravelbourg Formation; however, only one well is presently producing from this unit.

The dissolution of salt beds from the Middle Devonian Prairie Formation appears to have played a significant role in the depositional and hydrocarbon accumulation history of the Wapella-Moosomin area.

Kreis, L.K., Gent, M.R. and Vigrass, L.W. (1991): Subsurface brines in southern Saskatchewan; in Christopher, J.C. and Haidl, F.M. (eds.), Sixth International Williston Basin Symposium, Sask. Geol. Soc., Spec. Publ. No. 11, p283-292.

Subsurface brines with anomalously high calcium and magnesium content are widespread in reservoirs of Cambrian to Devonian age in southern Saskatchewan. Geographically, the anomalous brines are reported most commonly in the potash mining area around Saskatoon (lat. 52° N, long. 106° 30' W) and in the southeastern part of the province (lat. 49° 30' N, long. 103° W). Some of the brines occur as potash-mine inflow and these have been shown to have anomalously high contents of bromine, iodine and strontium. Origin of the brines is uncertain but they appear to have undergone a complex evolutionary history that may have involved dissolution of evaporite minerals in the Prairie Formation.

Data on characteristics of the reservoirs containing the brines are limited. Analysis of mechanical logs and a few drill-stem tests show that some of the reservoirs have significant thicknesses and permeabilities, and at least one may be large enough for commercial development.

In the Michigan Basin, subsurface brines provide the raw material for a billion dollar chemical industry. Preliminary work indicates that some of the Saskatchewan brines have calcium, magnesium, bromine and iodine contents that compare favourably with concentrations of these components in the commercial brines of the Michigan Basin.

Lewry, J.F., Thomas, D.J., Macdonald, R. and Chiarenzelli, J. (1990): Structural relations in accreted terranes of the Trans-Hudson Orogen, Saskatchewan: Telescoping in a collisional regime; in Lewry, J.F. and Stauffer, M.R. (eds.), The Early Proterozoic Trans-Hudson Orogen of North America, Geol. Assoc. Can., Spec. Pap. 37, p75-94.

The Reindeer Zone, western Trans-Hudson Orogen, is mainly a collage of juvenile, accreted(?) Early Proterozoic oceanic arcs and arc-related terranes. Structural relations and strain variation in the Saskatchewan part of the zone suggest that the main tectonism involved extreme crustal shortening, fold/thrust stacking, and the transportation of major allochthonous sheets, some evidently soled by initially flat-lying, regionally extensive high-strain zones. These early structures are refolded about more upright north-trending folds and later, regionally more important, northeast-trending folds to produce the present gross lithostructural patterns. Early structures were modified and, in many places, oversteepened during this refolding and also disrupted by late faulting. Modification of the prior domainal subdivision of the Reindeer Zone, involving revised definition and nomenclature of lithostructural elements, is proposed to more accurately reflect gross "three-dimensional" structural geometry.

La Ronge Domain is a stack of variably strained northwest/north-dipping tectonostratigraphic packages, most of which are bounded by high-strain zones. Glennie Domain largely comprises refolded nappe sheets soled by early mylonitic porphyroclastic gneisses/schists (Guncoat and Nistowiak gneisses). Early folds in much of Kisseynew Domain were probably initially highly overturned to recumbent. Flin Flon Domain is viewed as an imbricate thrust wedge soled by the Sturgeon Weir thrust and overlain by highgrade nappes of the Kisseynew Domain.

Windows of Archean basement are exposed at the lowest structural levels in the Glennie Domain and Hanson Lake Block, below allochthonous Early Proterozoic nappe sheets and their high-strain thrust soles. Thus, it seems possible that the Reindeer Zone is partly or wholly underthrust by lower plate continental lithosphere emplaced during terminal continental collision subsequent to arc formation. Seismic investigation and dating/isotopic characterization of late-/post-kinematic granites could test this possibility.

O'Hanley, D.S., Kyser, T.K., Rees, M.I. and Sibbald, T.I.I. (1991): The age of the auriferous granites and mesothermal gold mineralization, Goldfields area, Saskatchewan; GAC/MAC Joint Annual Meeting, May 1991, Toronto, Prog. Abstr., v16, pA92.

Previous U-Pb zircon studies of the North Shore Plutons of the Beaverlodge area in northern Saskatchewan yielded a spectrum of ages from 2356 ± 14 Ma (MacIntosh Granite) to 1994 ± 37 Ma (Box Mine Granite). This range of ages is in contrast to field relations which indicate that the granites, some of which host lode-gold quartz veins, were temporally related during emplacement (or formation) within the dominantly metasedimentary Murmac Bay Group. Single zircon ages obtained from the Mine Granites, using the Pb-evaporation technique, indicate that the age span of 360 Ma obtained from previous U-Pb zircon studies can be attributed to mixing of inherited zircon with primary zircon formed during the crystallization of the granites. The inherited zircons consist of a population having ages of near 2900 Ma and another having ages near 2500 Ma. The zircon populations are derived from proximal basement rocks with U-Pb ages of ≈3.0 to 2.9 Ga and ≈2.6 to 2.5 Ga. Zircons having the lowest $^{207}Pb/^{206}Pb$ ratios indicate that the Mine Granites were emplaced at ≈2000 Ma. The span of Pb-Pb ages from zircons in the Mine Granites is consistent with incorporation, by the granites, of detrital zircons from the quartzites of the Murmac Bay Group, into which the granites were emplaced.

Rb-Sr ages of gangue minerals in lode-gold quartz veins within the Mine Granites suggest that the deposits formed ≈1840 to 1900 Ma. Initial $^{87}Sr/^{86}Sr$ ratios of the gangue minerals are different for each granite, but are all greater than 0.757. These high and distinct initial Sr isotopic compositions indicate that old and discrete reservoirs were tapped by the mineralizing fluids.

Sibbald, T.I.I., O'Hanley, D.S., Kyser, K., Rees, M.I., Appleyard, E.C. and Roberts, R.G. (1991): Geology of the gold deposits, Goldfields, Saskatchewan; The Geological Society of CIM, First Annual Field Conference, Sept. 1991, Saskatoon, Prog. Abstr., p14-15.

The Goldfields gold deposits north of Lake Athabasca are hosted by granitic rocks (mine granites) emplaced within a shelf-type supracrustal sequence of Lower Proterozoic age. The supracrustals comprise quartzites, quartz-mica schists, dolomitic marbles, diopsidic calc-silicates and cordierite-anthophyllite-biotite schists. Amphibolites of predominantly mafic composition and of presumed igneous origin form sill-like units up to a few tens of metres thick.

Granitic rocks occur both as a basement to the supracrustals (Lodge Bay Granite) and as a granitic complex emplaced within the supracrustals. The granitic complex, which includes the mine granites, comprises several intermixed units with transitional relationships, both with each other and with the supracrustal rocks. Both magmatic and metasomatic processes are thought to have been instrumental in its formation.

The area is characterized by two major episodes of deformation. The first gave rise to the dominant regional foliation and the second to the major Goldfields syncline, whose axis plunges gently south and whose axial surface is subvertical and northerly trending. Regionally, both episodes appear to be characterized by inhomogeneous strain resulting in high and low strain zones and occasionally in discrete shear zones. Peak lower amphibolite facies metamorphism, marked by major new crystal growth and recrystallization, occurred after the first deformation. Greenschist facies retrogressive metamorphism accompanied the second deformation episode.

The mine granites comprise lenticular sheets of leucocratic, homogeneous, felsic rock typified by disseminated pyrite and cut by stockworks of gold-bearing quartz veins. The Box and Athona granites contain distinct, 0.5 cm diameter, euhedral quartz, K-feldspar (microcline-perthite) and albite megacrysts in contrast to the Frontier granite which is more nearly equigranular. Post-crystalline second deformation cataclasis is variable, being most marked in the Athona and least apparent in the Frontier.

Chemically the Box and Athona granites are peraluminous alkali feldspar, S-type granites, whereas the Frontier is a quartz rich (>80% SiO_2) and albite poor (<0.5% Na_2O) granitoid.

Structural investigations, restricted to the Box and Athona deposits, have shown three groups of quartz veins of which Group 2 are significantly auriferous. Older but related Group 1 veins parallel D_1 foliation, and Group 3 veins are subvertical easterly-trending, vuggy, breccia veins of much younger age. Group 2 veins occur in five orientations, although not all of these are represented in both mine granites. In the Box granite the veins occupy brittle fractures, whereas in the Athona granite they parallel brittle ductile shears. These fractures and shears are found to belong to a common strain regime, itself related to the geometry of the D_2 fold structures (Goldfields syncline).

Fluid inclusions in quartz from the auriferous veins in the Box, Athona and Frontier granites resemble those from mesothermal Archean lode gold deposits. They comprise primary inclusions, associated with the gold mineralizing event, and secondary inclusions reflecting the access of later fluids. Primary inclusions are characterized by a high density H_2O-CO_2 fluid, with low salinity (6 wt % NaCl equiv.) and median homogenization temperatures of 250°C to 300°C. CO_2 content varies from about 5 to 90 volume %, but tends to be either high or low, indicative of fluid immiscibility and thereby a possible deposition mechanism, although there are signs of inclusion necking. Inclusion data suggest minimum trapping pressures of 2.5 kb to 3.0 kb.

Stable isotopic values of vein quartz, feldspar, chlorite and tourmaline also resemble those of Archean mesothermal lode gold deposits and calculated vein water composition ($\delta^{18}O = 8.6$, $\delta D = -30$) is compatible with that of metamorphic water. Calculated equilibrium temperatures from quartz-chlorite and quartz-tourmaline mineral pairs are similar at around 450°C, and are thought to be the temperature of vein formation, whereas those of quartz-feldspar pairs from the host granites are typically variable suggesting disequilibrium due to later isotopic exchange.

U-Pb zircon dating of the mine granites and nearby plutons of geologically similar field age has yielded a spectrum of dates between 1977 ±37 Ma (Box granite) and 2356 ±14 Ma (MacIntosh Bay granite). Recent Pb-Pb single zircon investigations suggest that this age span of 360 Ma arises through mixing of inherited zircons with those formed during granite emplacement. The inherited zircons consist of populations with ages ca. 3.0 Ga and 2.5 Ga, which are characteristic of basement rocks in the Greater Beaverlodge area. Zircons with the lowest $^{207}Pb/^{206}Pb$ ratios suggest that the mine granites were emplaced at ca. 2000 Ma. This date is in agreement with earlier Rb/Sr whole rock isochron determinations from the same intrusive bodies. Rb/Sr ages of gangue minerals in the gold-quartz veins yield ages in the range 1840 Ma to 1900 Ma and exhibit high (>0.757) and individually distinct initial $^{87}Sr/^{86}Sr$ values suggesting that the gold mineralizing fluids tapped discrete reservoirs in old basement rocks.

Multidisciplinary studies at Goldfields outline a coherent geological history for the area and its contained gold deposits. Lower Proterozoic shelf-type supracrustals deposited on a ca. 3.0 Ga granitic basement were intruded and metasomatically replaced by synorogenic granitic fluids some 2000 Ma ago. This event was coeval with regional subduction and continental collision in the Taltson magmatic zone to the northwest. Gold mineralization occurred some 100 Ma to 200 Ma later during regional deformation, accompanied by retrogressive greenschist facies metamorphism, which led to development of northerly trending upright folds. Mesothermal auriferous metamorphic fluids, tapped from reservoirs in the older basement were introduced into brittle and brittle-ductile structural traps.

Thomas, D.J. and Harper, C.T. (1991): Gold in northern Saskatchewan: Alternative metallogenic models; The Geological Society of CIM, First Annual Field Conference, Sept. 1991, Saskatoon, Prog. Abstr., p20-21.

This paper reviews the diversity of geological settings in which gold occurs in northern Saskatchewan and draws attention to potential metallogenic environments. The majority of gold occurrences are in the greenstone-plutonic belts of the La Ronge, Flin Flon and Glennie Domains. As most of the discoveries are classified as structurally-controlled or shear-hosted lode gold mineralization, there has been a strong bias toward this style of mineralization in exploration strategies in recent years. Significantly fewer gold occurrences are reported from metasedimentary-dominated or high grade gneissic terrains. Is the relative abundance of gold occurrences in the greenstone belts a function of more favourable metallogeny, or is it simply an exploration bias toward greenstone belts and shear-hosted mineralization?

This paper will emphasize the other widely differing geologic and metallogenic environments in which economic and subeconomic gold concentrations are found. Most of these environments have received only a cursory level of attention and include: (1) gold mineralization in high grade gneisses of the Mudjatik Domain (Ithingo Lake); (2) stratabound veins in meta-arkosic rocks (North Lake); (3) iron formation-associated gold (Wedge Lake); (4) metagreywacke-hosted quartz-gold veins (Pine Lake showing); (5) epithermal Au-Ag mineralization associated with felsic volcanism (Laurel Lake); (6) Cu-Au-Mo-W porphyry-associated mineralization (Phantom Lake); and (7) gold mineralization related to S-type granites (Box, Athona mines). Although most of these occurrences display some degree of early mineralization or gold enrichment, later metamorphism and deformation have generally had a major role in formatting the final nature of the mineralization. Gold mineralization in older terrains can commonly be demonstrated to be the product of multistage geological processes involving the cumulative effects of meteoric, hydrothermal, magmatic, and metamorphic fluids over time.

Shear-hosted lode gold mineralization in the major greenstone belts will continue to be the "model of choice" in many prospecting programs. However, the creative application of various gold metallogenic models may lead to new deposits in previously neglected areas of northern Saskatchewan.

Thomas, D.J., Reilly, B. and Watters, B. (1991): Tectonostratigraphic controls on massive sulphide deposits in the Amisk Group: Flin Flon-Amisk Lake area; The Geological Society of CIM, First Annual Field Conference, Sept. 1991, Saskatoon, Prog. Abstr., p26.

The Flin Flon-Snow Lake belt is an easterly-trending Early Proterozoic greenstone belt extending from Hanson Lake, Saskatchewan to Snow Lake, Manitoba along the southern edge of the Precambrian Shield. The belt comprises an older volcanic-sedimentary succession (Amisk Group, >ca. 1886 Ma) which is unconformably overlain by molasse sediments (Missi Group, ca. 1840), and intruded by major subvolcanic (1892 to 1876 Ma) and post-volcanic (1860 to 1840 Ma) plutons as well as post-Missi intrusions (ca. 1837 to 1773 Ma). Thermotectonism accompanied the on-set of post-volcanic plutonism (ca. 1860 Ma) and terminated ca. 1800 to 1780 Ma.

In the Flin Flon-Amisk Lake area the Amisk Group comprises a collage of contrasting volcanic regimes. Chemical and physical characteristics of the volcanic sequences indicate a variety of environments including: a mature calc-alkaline island arc setting comprising intermediate to felsic subaqueous to subaerial rocks intruded by high level subvolcanic felsic intrusions (west Amisk Lake); mafic subaqueous flows of tholeiite, ocean floor or back-arc spreading affinity (east Amisk Lake and West Arm, Schist Lake); and mafic flows and tuffs of tholeiitic to calc-alkaline, oceanic island arc affinity (Flin Flon-Phantom Lake). Preliminary investigations indicate these contrasting volcanic environments are tectonically discordant bounded by zones of brittle-ductile shearing and alteration which predate deposition of the Missi Group and emplacement of some of the major plutons. This structural event is tentatively interpreted as a period of early northeast-directed thrusting within the Amisk Group. This is best exemplified in the West Arm (Schist Lake) area where an oceanic island arc is juxtaposed with a possible back arc spreading environment.

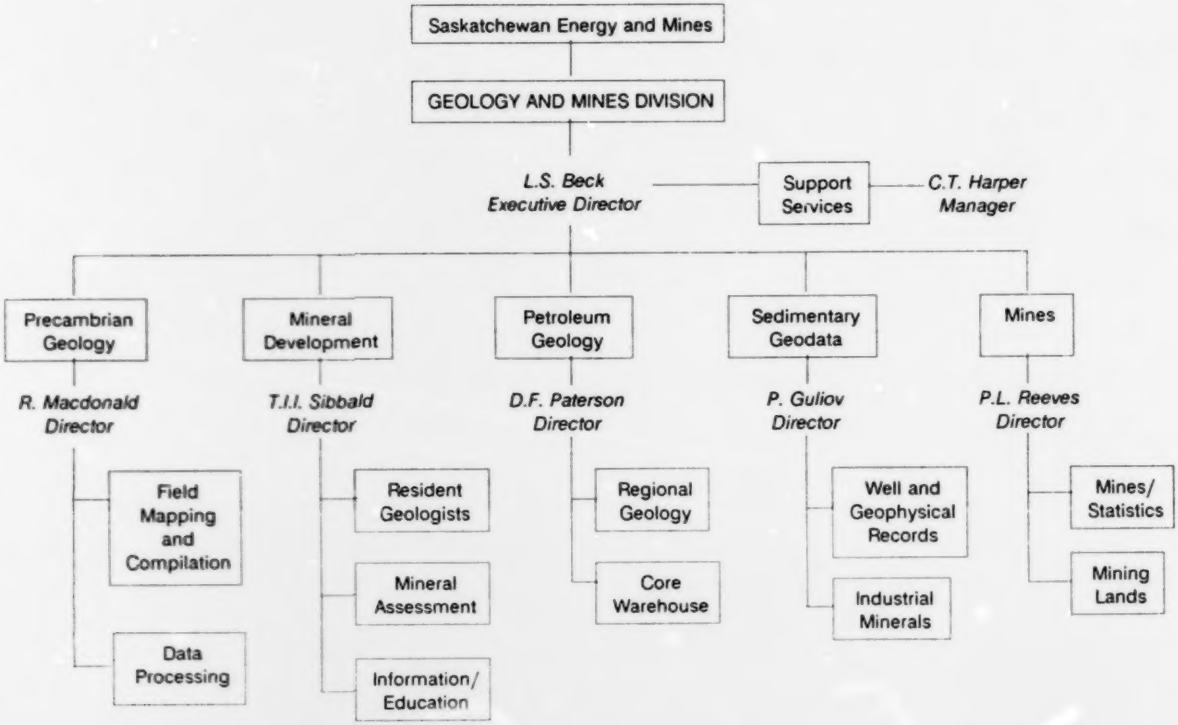
Thirty-three major massive sulphide deposits have been discovered to date in the belt (cumulative 150 million tonnes) of which 22 have had commercial production. The largest deposit, the Flin Flon Mine (HBMS), began production in 1930 and has produced about 63 million tonnes of ore grading 2.17% Cu and 4.1% Zn. Although the deposits have typical VMS characteristics, they vary greatly in ore grade, tonnage, host rocks and alteration.

The premise of this paper is that the distribution and ore tenor of massive sulphide deposits in the Flin Flon-Amisk Lake area reflect inherent differences in the volcanotectonic environments represented in the Amisk Group. Early thrust imbrication juxtaposed volcanic regimes of contrasting plate tectonic affinities. The differences in grade, size and mineralogy of base metal deposits in this area can be assessed in light of this new tectonostratigraphic model.

Acknowledgments

Saskatchewan Geological Survey project leaders acknowledge the assistance in the field and office of Margaret Bertoldi, Darren Burrows, Trevor Good, Kevin Halabuza, Dale Heibein, Christie Langen, Wayne Laturnas, Alain Leclair, Daniel L'Heureux, Mark Malfair, Glen McCrimmon, Dwayne Morrison, Erik Nickel, Ian Russel, Tom Schmidt, Donna Schreiner, Marilyn Schultz, Mark Sombach, Jennifer Stack, Jason Sturgeon, Terri Uhrich, Andrew Wallace, Donald Williams and Adam Zasada.

The editors are grateful for assistance given in the editing and production process by Laura Smith, Judy Adams and Horst Stolz. This report was prepared with Ventura Desk Top Publisher and laser printer. This year for the first time many of the manuscript maps were compiled using AutoCAD/FieldLog software; Brent Czornobay played a critical role in facilitating this process. Phil Weir handled preparation of some of the figures, and with assistance from Jim Bedo, Rick Pearce and Harry Buyze, assembled the various components of the maps to the plate-ready negative stage.


Saskatchewan Energy and Mines
GEOLOGY AND MINES DIVISION
L.S. Beck
Executive Director
Support Services
C.T. Harper
Manager
Precambrian Geology
R. Macdonald
Director
Field Mapping and Compilation
Data Processing
Mineral Development
T.I.I. Sibbald
Director
Resident Geologists
Mineral Assessment
Information/ Education
Petroleum Geology
D.F. Paterson
Director
Regional Geology
Core Warehouse
Sedimentary Geodata
P. Guliov
Director
Well and Geophysical Records
Industrial Minerals
Mines
P.L. Reeves
Director
Mines/ Statistics
Mining Lands

The Saskatchewan Geological Survey comprises the geoscience activities of the Precambrian Geology, Mineral Development, Petroleum Geology and Sedimentary Geodata Branches, and is represented by R. Macdonald, Chief Geologist.

Staff as of November 1, 1991

At 1914 Hamilton Street, Regina:

Geology and Mines Division Administration

Executive Director	L.S. Beck, *Ph.D, P.Eng.*
Secretaries	J. Adams
	V. Ziegler

Saskatchewan Geological Survey

Chief Geologist	R. Macdonald, *Ph.D, P.Eng.*

At 1914 Hamilton Street, Regina:

Precambrian Geology Branch

Director	R. Macdonald, *Ph.D, P.Eng.*
Field Mapping and Compilation:	
Project Geologists	G.D. Delaney, *Ph.D*
	K.E. Ashton, *Ph.D*
	B.A. Reilly, *M.Sc.*[1]
	W.L. Slimmon, *B.Sc.*[1]
Geologist-geophysicist	B. Czornobay, *B.Sc.*[1]
Data Processing:	
Geologist	N. Lee, *M.Sc.*[3]
Secretary	J. Bilkhu[2]
Warehouse Operator (part-time)	E. Gaw

Mineral Development Branch

Director	T.I.I. Sibbald, *Ph.D, D.I.C.*
Mineral Deposit Studies/Compilation/Education:	
Geologists	D.G. MacDougall, *B.Sc.(Hons.)*
	D.J. Thomas, *M.Sc.*
	Vacant
Mineral Assessment:	
Geologists	H. Stolz, *B.Sc.*
	R.W. Bennett, *B.Sc.(Hons.), P.Eng.*
Geological Assistant	L. Smith, *B.Sc.*
Resident Geologist Offices:	
La Ronge:	
Geologists	A.J. Gracie, *Ph.D, P.Eng.*
	P.L. Schwann, *M.Sc.*
District Mineral Officer	W. Hounjet
Secretary	S. Robinson
Creighton:	
Secretary (part-time)	L. Skipper

At 1914 Hamilton Street, Regina:

Sedimentary Geodata Branch

Director	P. Guliov, *M.Sc., P.Eng.*
Industrial Minerals:	
Geologists	M.R. Gent, *M.Sc.*
	F. Swanson, *B.Sc.*[1]
Well and Geophysical Records:	
Geologist	D.R. Troyer, *M.Sc., P.Eng.*
Coding Clerk	R. Kovacevic
Well File Clerk	J. Tanouye
Geophysical Clerk	K. Barrett

At 201 Dewdney Avenue East, Regina:

Petroleum Geology Branch

Director	D.F. Paterson, *Ph.D, P.Eng.*
Regional Geology:	
Geologists	C.F. Gilboy, *Ph.D, P.Eng.*
	F.M. Haidl, *M.Sc.*
	L.K. Kreis, *M.Sc.*
Warehouse:	
Supervisor	D. Forsberg
Staff	R.L. McCall
	R. Lavigne
	D. Gallant
Secretaries	M. McNally
	L. Bleau

At 1914 Hamilton Street, Regina:

Support Services Section

Manager	C.T. Harper, *Ph.D, P.Eng.*
Review Geologist	C.L. Brown, *B.Sc.*[1]
Geological Survey Drafting Unit:	
Supervisor	R.P. Kydd
Draftspersons	P.A. Weir
	K.G. Jones
General Drafting Unit:	
Supervisor	J.J. Bedo
Draftspersons	R.O. Pearce
	A.H. Buyze

---

[1] Position contracted
[2] Also serving Mineral Development Branch
[3] Seconded to Mines Branch

MAPS
NOT
FILMED

CARTES
OMISES